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PRECAMBRIAN GEOCHRONOLOGY ON THE ABSOLUTE-TIME SCALE[1]

by

N. P. Semenenko

In the light of much accumulated data on the absolute age of Precambrian formations, it became possible to formulate a new approach to the correlation of Precambrian sections and to determine the main stages in the formation of the earth's sial crust.

The absolute-age figures reveal the formation time of crystalline lattices in minerals whose radioactive decay products determine the absolute age. They reflect the age of the mineralization, metamorphism, and igneous activity. The mineralization time corresponds to that of the folding and of the transformation of a mobile geosynclinal folded province into a platform. The absolute-age figures for extrusives, included in sedimentary-extrusive sequences, as well as those for glauconite from nonmetamorphosed deposits, reflect the formation time for the deposits themselves.

In a paper read at the Sixth Session of the Commission for the Absolute-Age Determination of geologic formations in 1958, the author reviewed the Precambrian sections in the light of new data on the absolute age for Ukrainian and Baltic shields, Russian platform, Urals, and the shields in Siberia, Canada, and Africa.

Within the small space of a magazine article, we shall only touch upon the outstanding data used in the absolute-age determination of Precambrian formations.

Extensive factual material on the Ukrainian Precambrian has been gathered as a result of the absolute-age determination by the uranium-thorium-lead method, at the Institute of Chemistry and Analytical Chemistry, Academy of Sciences, U.S.S.R., by A.P. Vinogradov and A.I. Tugarinov; at the Radio Institute; Academy of Sciences, U.S.S.R., by L.V. Komlev; at the Institute of Geological Sciences, Academy of Sciences, U.S.S.R., by N.P. Semenenko, Ye.S. Burkser and M.N. Ivantishin; and in other laboratories. On the basis of these data, we have separated seven age groups (reading downward) as follows, in million years:

1. The Azov alkaline complex, 900 to 1000 (?).

2. The Korosten complex (the Korsun'-Novomirgorod and Uman' massifs), 1100 to 1250.

3. The Ovruch sedimentary volcanic metamorphic series and granites of the Perzhansk and Osnitsk massifs, 1300 to 1400; granites of the Bokovyansk, Verblyuzhiy, and other massifs in the central part of the Ukrainian crystalline shield, 1300 to 1600.

4. The Saksagan and metabasic series and granites of the Ingulets, Bobrinets, and Mokromoskovskiy massifs, 1700 to 1820.

5. The Pobuzh (Bug) series of pyroxene-plagioclase and other gneisses and charnockite-monzonites, 1900 to 2000.

6. The Podol'sk gneiss series and the Chudhov-Berdichev granites, 2000 to 2250.

7. The Dnieper series of gneisses and granites, 2300 to 2600 and older.

Sedimentary-volcanic formations, younger than 1300 to 1400 million years, are lacking in the Ukrainina Precambrian; only in the west are there Riphean platform sedimentary formations, 510 to 560 million years old, overlying with a major break the Podol'sk shield crystallines.

In Karelia, in the eastern part of the Baltic shield, A.A. Polkanov has identified four groups of granites corresponding to the four times of folding. According to E.K. Gerling's determination of the micas' age by the potassium-argon method, their age is as follows, in millions of years:

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[1]Geokhronologiya dokembriya v absolyutnom letoischislenii.

1. Jotnian and the fourth granite group, 1500;

2. Karelian and the third granite group, 1560;

3. Bothnian and the second granite group, 1800;

4. Belomorian and the first granite group, 1850 to 1900.

K.O. Krats and L.Ya. Kharitonov combine the Bothnian and Karelian folding times into a single Karelian time, subdividing its rocks into three structural stages. The metamorphism age of the lower structural-stage schists is presumed to be 1740 million years.

According to recent data of E.K. Gerling and A.A. Polkanov, as reported at the Sixth Session of the Commission on Absolute Age, rock ages as much as 3480 million years in the western part of the Baltic shield are given for micas, by the potassium-argon method. The ancient rocks, 2000 to 3400 million years old, have not been differentiated areally, as yet, and are assigned by A.A. Polkanov to the Saamidian and Catarchaean. On the basis of these new data, E.K. Gerling has recognized the following seven age groups, in millions of years: 1 -- 3060 to 3480, 2 -- 2530 to 2880, 3 -- 2000 to 2450, 4 -- 1880 to 2000, 5 -- 1710 to 1830, 6 -- 1530 to 1700, and 7 -- 1010 to 1180.

In the Baltic shield, in Sweden, the following series have been differentiated by L. Magnuson in millions of years:

1. Algonkian, older than 500;

2. Dalaslandian, 900 to 1050 (1000);

3. Gothian and Karelian; the age relationship between them has not been established, but they are younger than the underlying Svionian.

4. Svionian, 1800.

L. Magnuson believes that the pre-Gothian series is older than the Svionian.

The Dalaslandian and Svionian age have been dated by the uranium-thorium-lead method as follows: for Dalaslandian metamorphic schists, 1000 million years; with 1700 to 1760 for the Svionian. This age has been confirmed by the potassium-argon method, at the Geology Institute, Academy of Sciences, Ukr.S.S.R.

An age of 1050 million years has been determined for the Arendale cleveite, southern Norway. According to Holmes, the Norwegian breggerites from Moss are 850 to 905 million years old. The upper parts of the Precambrian section, younger than those of the eastern part of the Baltic shield in the U.S.S.R., are developed in Sweden and Norway, to the west.

A number of Precambrian age groups has been established from isolated data, on the Russian platform and the Urals:

1. Riphean platform deposits are 680 to 690 million years old, as determined by N.I. Polevaya (from glauconite from the Serdob borehole); basalts from the Podol'sk-shield Riphean is as much as 570 million years old, as determined by the Geology Institute, Academy of Sciences, Ukr.S.S.R.

2. The Taratash formation and the Berdyaush massif in the Urals is 950 to 1080 million years old, as determined by L.N. Ovchinnikov, by the potassium-argon method.

3. The Tatar Arch formation of the second Baku area, Bashkiria, is 1500 to 1600 million years old, according to M.I. Garris.

4. The Kursk series of metamorphic and crystalline rocks is 1740 million years old, as determined by the potassium-argon method, at the Geology Institute, Academy of Sciences, Ukr.S.S.R.

The age of older series, underlying the Kursk ferruginous series, has not been determined.

A number of the absolute-age determinations has been made for the Precambrian of Siberia (Aldan and the Sayan-Baikal highlands): by E.K. Gerling, at the Precambrian Geology Institute, Academy of Sciences, Ukr.S.S.R. and by N.I. Polevaya, at the All-Union Geological Institute (published in papers of D.A. Velikoslavinskiy and S.V. Obruchev). A number of determinations for metamorphic schists of the Chuya formation and the Mama Bodaybo series have been performed by the potassium-argon method, at the Geology Institute, Ukr. S.S.R. The Ages of orthite, monazite, and betafite from the Slyudyanka River area have been determined by the uranium-thorium-lead method, at the V.I. Vernadskiy Institute of Geochemistry and Analytical chemistry, Academy of Sciences, U.S.S.R.

We have combined all of these known absolute-age determinations into four Precambrian age groups, ranging from 500 to 1900 million years old, in Table 1.

According to these data, the Baikal folded zone was developed for 500 to 1000 million years as a geosynclinal province, and is made up chiefly of Riphean rocks. As noted by M. M. Odintsov, N.A. Florensov, and P.M.

Table 1

Determination of the absolute age of geologic formations from the eastern Siberia regions.

	Locality and complex	Mineral and rock	Absolute age in million years	Age group in million years
Paleozoic	Mama granite field along the Kunkudera River	Granites	250-280	
	Mama belt, Mama granites	Muscovite from pegmatite	300-360	
	Eastern Sayans, Biryusa River, pegmatites	Muscovite from pegmatites	488-478	460-490
	Ust'-Mama district, Mama sequence	Pegmatites	460	
Precambrian	Slyudyanka River, Lake Baikal	Orthite, monazite, betafite	500-600	500-700
	Chuya granite field (Mama-Chuya district)	Granites and granite-gneisses	450-760	
	Bodaybo (green) series	Phyllite schists	500-550	
	The Mama-Olongro and Mama-Dovgokit interfluve area	Knotty phyllite schists	670-700	
	Mama district, Vitim River	Schistose pegmatite	700	
	Chuya granite field, Mama-Chuya district	Granite	970-1000	960-1000
	Mama-Chuya district, mouth of Chuya Bram'ya	Green phyllite schists of the Chuya formation	970	
	Proterozoic gneisses from the southern rim of the Aldan shield	Gneisses	960	
	Kalar massif	Granites	980-1000	
	Aldan, Nemkyr district	Coarse-grained granite	1300	1300-1580
	Same locality	Alaskite	1580	
	Eastern Sayan, Oka River	Muscovite	1670	1670-1900
	Aldan, Taiga district	Phlogopite	1780-1900	

Khrenov, the geosynclinal conditions of that province changed at the close of the Riphean to transitional subgeosynclinal conditions of the lower Paleozoic. Related to this Paleozoic igneous activity is the intrusion of plutonic granitoids and the formation of muscovite pegmatites of the Biryusa and Mama districts, whose micas have been determined to be 460 to 490 and 300 to 360 million years old, respectively. The age of the Mama granite field along the Kunkudera River has been determined as 250 to 280 million years, which suggests a prolonged igneous activity in that region, persisting as late as the late Paleozoic.

Metamorphism of both the Baltic complex and the Bodaybo and Mama series lasted for 500 to 700 million years. The Chuya granites, like the gneisses from the southern rim of the Aldan shield and the Kalar massif granites, are 970 to 1000 million years old.

More ancient age groups, 1300 to 1600 and 1700 to 1900 million years old, are present in the Aldan and are known from the ancient eastern Sayan block on the Oka River (1670 million years old). The age of the charnockite series in the Aldan has been determined by phlogopite, as 1900 million years.

In the Precambrian shield of Canada, the following age provinces have been differentiated, according to the absolute-age data of D. Wilson, R.D. Russel, R.M. Farker, D. Davis, G.R. Tilton, and others (in million years):

Keewatin, (Manitoba) 2600;
Yellowknife, 2200 to 2400;
Athabaska (Saskatchewan), 1800 to 1900;
Great Bear Lake, (Gunidon, Colorado) 1405 to 1440;
Grenville, 880 to 1140.

Finally, the oldest-age determinations for the Appalachians is 690 million years, obviously for the uppermost Precambrian.

In Africa, eleven age groups have been recognized, according to L. Cahen [21]: 1 -- younger than 485; 2 -- 485 to 600, assigned to the Eocambrian cycle; 3 -- 600 to 700, the Katanga cycle; 4 -- 800 to 1200, Gordonia cycle; 5 -- 1200 to 1400, Karagwe-Angola; 6 -- 1650 to 1850, Uganda; 7 -- 2000, Witwatersrand; 8 -- 2000 to 2300, Limpopo; 9 -- 2650, Kibali-Shamvaian; 10 -- 2900, (here belong Nyasaland and Sierra Leone); 11 -- 3000 to 3800 (Dodoma, Sebaquian, Swaziland).

We have compiled a composite Precambrian chronologic table for various parts of Africa and Madagascar (Table 2).

Although these data of the absolute-time determination are of various degrees of reliability and precision, they make it possible to outline a correlation outline for Precambrian formations and to posit the problem of major subdivisions of Precambrian geologic history, embracing the entire earth.

A correlation for the absolute age of minerals and rocks is drawn by establishing common data for mineralization cycles associated with Precambrian folding epochs. Ten such cycles have been designated (figures in million years).

Cycle One, 500 or 520 to 700, corresponds to the Baikal folding epoch, well developed also in the Yenisey Range and the Sayan-Baikal Arch. The associated radioactive minerals of the Slyudyanka complex are 500 to 600 million years old; those of the Mama series, 700.

Corresponding to the Baikal folding epoch on the Russian platform are Riphean deposits, 520-570-690, and the Timan folding epoch. In Africa, the 630 age is assigned to the first cycle (Katangian) which is developed in the Katanga geosyncline, South Africa (Waterberg system), and elsewhere. The same age of mineralization (690) has been determined from lower Appalachian structural stages. The Algonkian series of Sweden appears to belong here.

Cycle Two, 800 to 1140 or 1200. Here belong the Chuya folded belt of the Siberian Baikal Arch, with the Chuya granites and metamorphic schists of the Chuya River, 970; gneisses from the south rim of the Aldan, 960; the Taratash zone in the Urals, with intrusions of the Berdyaush granites, 950 to 1000; the Tatar Arch gabbro-diabase intrusion, on the Russina platform, 1140; and the Azov intrusives of the Ukrainian shield.

In Africa, this cycle includes the Gordonian mineralization, 1025, also including the Transvaal system and others; the Madagascar Chapolina system, 1120; in Baltic shield it also includes the Arendal mineralization, 1000; the Norwegian breggerites, 850 to 900; the Dalaslandian series of Sweden, 900 to 1100 or 1050; and the Grenville mineralization of the Canadian shield, 880 to 1140.

Cycle Three, 1150 to 1300, embraces the Karagui-Angola mineralization, 1200, of the African shield; the Kibari-Urundi cycle of Equatorial Africa and the Belgian Congo, 1300; and the Korosten' intrusive cycle in the Ukrainian shield, 1100 to 1250.

Cycle Four, 1300 to 1600: the Jotnian of the Baltic shield, accompanied by granite intrusions of group four, 1500; the Volynsk folding in the Ukrainian shield, with the deposition of the Ovruch series, 1400, and the intrusion of the Bokovyan-Verblyuzhiy granites; on the Russiah platform, granite intrusions, 1500 to 1600, have been identified in the Second Baku formation, in the Tatar Arch region; granite intrusions of Aldan, Siberia, 1300 to 1580; the Great Bear Lake mineralization of the Canadian shield, 1440.

Cycle Five, 1650 to 1850: the Russian platform and the Ukrainian shield witnessed the Saksagan folding accompanied by granite intrusions, 1700 to 1820, with the formation of the ferruginous-siliceous-schist-volcanic formations of the metabasic and Saksagan series, and of the terrigenous flysch upper stage formations separated, from them, by an unconformity.

The Karelian folding epoch in the eastern U.S.S.R. part of the Baltic shield, corresponded to the 1710 to 1800 age group and to the second and third granite intrusion groups. Ferruginous-siliceous-volcanic formations were deposited: the volcanic spilite-keratophyric and terrigenous, of the lower, middle, and upper Karelian structural stages. In Sweden, the Sinian mineralization, 1800, probably belongs to this cycle. In Siberia, it witnessed the Aldan folding, 1780 to 1900, with the deposition of a number of sedimentary formations separated by unconformities, according to Yu. K. Dzevanovskiy.

Table 2

Composite data on the Precambrian geochronology of Africa
(figures are absolute age in millions of years)

Age		French West Africa	North Africa, Sahara, Morocco	Central Equatorial Africa and Belgian Congo	South Africa	Madagascar
Precambrian IV	630	Biem, Nigritian	Precambrian III (rhyolites), Morocco	Katanga, 630; Madagarasi; Ubangi-Landi; Marungu	Lutete; Umkondo; Waterberg, 630; Damara, 600; Otawi, 630	Post-Chipolini granites, 645-700
	800-1200	Falemian, Rokel	—	Lulua, 1070	Lomagundi Transvaal, 840-1200 Kheis Gordonia, 1025	Chipolino, 1120
Precambrian III	1200	—	—	Kibara-Urundi, 1300 Karagwe-Angola, 1200-1400 Lundi	—	—
	1650-1850	—	Precambrian I in the Anti-Atlas, 1650	Uganda, 1825-1890	Ventersdorp Pittsburg, 1650-1850	Post-Vogelberg granites, 1800
	2000	Tarquian, 1950	—	—	Witwatersrand, 1800-2000	Vogelberg, 2000
Precambrian II	2000-2300	Birimian, 2190	Farisian	Ruzizi, 2270	Limpopo, 2150	—
	2650	—	—	Kibali, 2650 Kalundwe	Shamvaian, 2650 Bulavian, 2825 (Bikita)	Graphite, 2400-2600
	2900	—	—	Nyanzian, 2850	Lead, So. Rhodesia, 2700-3050	—
Precambrian I	3000-3800	Dahomeyian	Saharian	Eastern Nile, Banzyville, Dodoma, 3250	Sebaquian, 3200 Swaziland, 3500	Androen

The Uganda cycle, 1650 to 1850, took place in Africa; and the Athabaska mineralization, 1800 to 1900, in the Canadian shield. The Huronian series, 1300 to 1800 is known from the latter. The rocks of that series appear to have been formed during the fourth and fifth cycles.

Cycle Six, 1900 to 2000: The Bug folded zone formed in the Ukrainian shield, accompanied by charnockite intrusions, 1900 to 2000. The Baltic shield witnessed the Belomoriye folding and the intrusion of the first granite group, 1850 to 1900, or else of the second age group -- 1880 to 200, according to E.K. Gerling. The lower structural stages of the Aldan sedimentary section probably also belong to this cycle (older than 1900).

In Africa, this mineralization age group is developed in the south, in the Witwatersrand system, 2000; and in the Tarquian system, 1950.

Cycle Seven, 2000 to 2300: the Podol'sk folded zone of gneisses and the Chudnov-Berdichev granites, 2000 to 2250; the 2000 to 2450 age group of the western and the U.S.S.R. parts of the Baltic shield; the Limpopo mineralization cycle in the southern African shield, 2150; the Ruzizi system of Equatorial Africa, 2270; the Birimian folded system of West Africa, 2190, made up of sedimentary-volcanic and manganese-siliceous formations. The mineralization of the Madagascar Vogelberg' system of sedimentary-volcanic formations (granulites, amphibolites, schists, and chipolino) is 2000 to 2140 million years old.

The Yellowknife mineralization, 2000 to 2400 occurred in the Canadian shield.

Cycle Eight, 2300 or 2400 to 2650, is represented by the Shamvayan (2650) and Kibali, 2650 mineralization cycles of the African shield. Equally old, 2400 to 2600 is the graphite system mineralization of Madagascar made up of di-micaceous kyanite and graphite gneisses, granulites, and migmatites. The Keewatin province mineralization, 2600, took place in the Canadian shield.

In the Ukrainian shield, this age cycle, 2300 to 2600 and older, includes individual blocks of the Dnieper gneiss series. In the western part of the Baltic shield and in the U.S.S.R., this cycle includes the 2000 to 2450 age group.

Cycle Nine, 2560 to 2900, is known from the African shield: the Nyanzian, 2850; the Bulavian (Bikita), 2850; Sierra Leone, 2930. In the western part of the Baltic shield, this cycle includes the 2530 to 2880 age group.

Cycle Ten, 3000 and older. Here belong the Dodoma, 3250; Sebaquian, 3200; Swaziland, 3500; and the Katarchaean group, 3060 to 3480, in the western part of the Baltic shield.

* * *

The ten epochs of folding and associated mineralization cycles comprise time intervals in the history of the sial envelope of the earth, from 500 to 3500 million years. Each of these cycles is commensurable with post-Cambrian folding epochs -- the Caledonian, Hercynian, and Alpine -- which have taken place in the last 500 million years.

The Precambrian folding epochs and associated mineralization cycles apparently form four major planetary megacycles in the Precambrian history of the earth's crust, commensurable with the post-Cambrian. On the whole, five megacycles representing major eras, including the post-Cambrian, have been recognized:

- Fifth post-Cambrian -- three epochs of folding;
- Fourth Precambrian -- two epochs of folding;
- Third Precambrian -- three epochs of folding;
- Second Precambrian -- three epochs of folding;
- First Precambrian -- two (?) epochs of folding.

The Fourth Precambrian megacycle, 500 to 1100 or 1200, consisted of two epochs: first, the Baikal folding and its contemporaneous Katangian of Africa, 500 to 700; the second -- the Chuysk folding, 800 to 1150.

This megacycle witnessed the folding of the Baikal-Sayan Arch, Timan-Ural belt (Taratash formation; Katanga geosyncline, Grenville province, Algonkian and Dalaslandian series of Sweden. Mobile zones, initiated in this cycle, continued in their development in the post-Cambrian of the Baikal-Sayan, Timan-Ural, and other belts.

Both the Russian and Ukrainian platforms remained in a platform state.

The Third Precambrian megacycle, 1200 to 1850, had the following subdivisions: the third -- Korosten' -- epoch of the Ukrainian shield, which was represented in the Russian platform by the terminal stages of mobile zones of the third megacycle, also represented in the Karagui-Angola and Kibara-Urundi, Africa; the fourth epoch -- the Jotnian of the Baltic shield and the Volyn' epoch of the Ukrainian shield, 1400 to 1600, with the folding of the second Baku granites and gneisses, 1500 to 1600; the fifth epoch of the Saksagan folding,

1700 to 1850, with the contemporaneous Karelian folding, probably contemporaneous Svionian folding of Sweden, 1800; the Aldan folding of Siberia, etc.

In this megacycle, the Saksagan folding was initiated and developed in the Ukrainian shield and the Kursk Oblast' (province). In the Baltic shield (U.S.S.R.), the folding affected the Karelian rocks.

In the Ukrainian shield, the Saksagan folding was initiated on a different plan, across the lower structural trend. This marked the beginning of the second Precambrian megacycle.

In the fourth Precambrian epoch, the deformation of the Ukrainian shield in Volyn' developed along a different structural design: a geosyncline was initiated here. In the Russian platform, this epoch was reflected in the Tatar Arch. Jotnian disturbances took place, as well, in the Baltic shield.

The Siberian platform, in the fourth and fifth epochs of folding, apparently witnessed the formation of a northwest trending Aldan folding.

The younger Baikal Arch, west of Aldan, was initiated on a different structural plan, in the fourth Precambrian megacycle.

In the African shield, there was a major (200 to 250) intervention of mineralization between the cycles of 1200 to 1400 and 1650 million years. The fifth mineralization cycle -- Uganda -- 1650 to 1850, was developed here contemporaneously with the Saksagan rocks of the Russian platform.

In the Canadian shield, the second Precambrian megacycle corresponds to the Great Bear Lake province mineralization, i.e., it obviously corresponds to the development of folded systems make up chiefly of the Huronian series, 1300 to 1800.

The Second Precambrian megacycle, 1900 to 2650, included the sixth, seventh, and eighth mineralization cycles and the corresponding folding epochs. They are reflected in the lower basement of the Russian platform as three structural stages of the three mineralization cycles.

The Ukrainian shield witnessed the development of the sixth Pobuzh (Bug) and the seventh Podol'sk folding epochs. These two stages have a common structural plan. Less known is the distribution of the Dnieper folding blocks of a deeper structural stage, 2300 to 2600.

Belomoriye rocks were developed in the Baltic shield, evidently as an upper structural stage of the third Precambrian cycle, along with the structural stage of the age groups, 2000 to 2450 and 2500 to 2700. In the Aldan, that part of the age group older than 1900 probably belongs to the lower megacycle. It is possible that these deeper stages are not exposed there.

In the Canadian shield, the Yellowknife mineralization, 2200 to 2400, has been assigned to the seventh cycle; with the Keewatin mineralization, 2600, assigned to the eighth cycle of the second Precambrian megacycle and corresponding to the Keewatin and Kuchiching sedimentary-volcanic formations.

In the African shield, the structural stages of all three cycles were formed in the second Precambrian megacycle. The folded Tarquian system of West Africa and the Witwatersrand in South Africa, were formed in the sixth cycle, 1900 to 2000. During the seventh cycle, 2000 to 2200 or 2300, the Birimian folded system was developed in West Africa. It was oriented submeridionally, with its trend inherited by the overlying Tarquian structural stage. The Limpopo structural stage was in progress in South Africa. The deeper structural stages were formed during the eighth mineralization cycle, 2400 to 2650, in zones of Kibali, Shamvayan, and in the Graphite system of Madagascar.

Structural stages dating back to the first Precambrian megacycle, older than 2650, are known from the U.S.S.R. (Baltic shield) and from South Africa. They occupy only a comparatively small area but are the deepest roots of the earth's sial envelope.

* * *

The duration of megacycles and epochs of folding is as follows (in millions of years): Fifth post-Cambrian, 500 to 600; fourth Precambrian, 600 to 650 (from 500 to 1150); third Precambrian, 650 to 700 (from 1200 to 1850); second Precambrian, 750 to 800 (from 1900 to 2650); first Precambrian, 850 (from 2650 to 3500). The folding epochs and the mineralization cycles lasted for 150 to 350 or 400 million years (Table 3).

It is possible that the first Precambrian period should be designated as Katarchaean; the second, as Archaean; the third, as Proterozoic; and the fourth, as Riphean or Baikalian, because the latter best represents the uppermost Precambrian section.

The question of the Precambrian order of nomenclature -- whether ascending or descending -- is now under consideration. Thus, the Tectonic Map Conference proposed the

Table 3

Geochronology of megacycles and zones of folding

Cycles	Age in million years		Duration in million years
	from	to	
Fourth Precambrian megacycle			
First cycle: Baikal, Katangian	500	700	200
Second cycle: Chuya, Taratash (Urals), Gordonian (Africa), Azov (Ukraine), Grenvillian (Canada), Chipolinian (Madagascar), Dalaslandian (Sweden).	800	1150 (1200)	350
Third Precambrian megacycle			
Third cycle: Korosten' (Ukraine), Karagui-Angola (Africa)	1150	1300 (1250)	150-200
Fourth cycle: Jotnian (Karelia), Volyn' (Ukraine), Great Bear Lake (Canada)	1300	1600	250-300
Fifth cycle: Saksagan (Ukraine), Karelia, Aldan (Siberia), Uganda (Africa), Athabaska (Canada), Svionian (Sweden)	1700	1850 (1900)	150-250
Second Precambrian megacycle			
Sixth cycle: Pobuzh (Ukraine), Belomorian (Karelia), Tarquian (Africa), Witwatersrand (Africa)	1900 (1850)	2000	100-150
Seventh cycle: Podol'sk (Ukraine), Ruzizi, Birminian (Africa), Yellowknife (Canada)	2000	2300	300
Eighth cycle: Dnieper (Ukraine), Keewatinian (Canada), Shamvay, Kibalian (Africa), Graphite (Madagascar)	2300	2650	350
First Precambrian megacycle			
Ninth cycle: Bulavian, Nyanzian, Sierra Leone (Africa)	2650	2900	350
Tenth cycle: Dodomian, Swaziland (Africa)	3000	3400	400

descending order: Precambrian A-I, B-II, C-III, D-IV. The corresponding Precambrian differentiation is as follows, in millions of years:

(Riphean)

Precambrian I:	500-1150 or 1200
First cycle:	500-700 or 800
Second cycle:	800-1150

(Proterozoic)

Precambrian II:	1200-1850 or 1900
Third cycle:	1150-1300
Fourth cycle:	1300-1600
Fifth cycle:	1700-1850

(Archaean)

Precambrian III:	1900-2650
Sixth cycle:	1900-2000
Seventh cycle:	2000-2250
Eighth cycle:	2300-2650

(Katarchaean)

Precambrian IV:	2650-3400
Ninth cycle:	2650-2900
Tenth cycle:	3000-3400

* * *

The division of rocks into two groups, the Proterozoic and Archaean, is more or less conditional for different platforms and shields. The decisive role in assigning widely separated rocks to either Archaean or Proterozoic was the degree of metamorphism, which cannot be the measure of time. The absolute-age determination, despite the many difficulties, puts Precambrian geology on a more substantial basis.

The terms, Proterozoic and Archaean, have lost their original meaning. For instance, the oldest rocks of the Ukrainian shield, as much as 2200 million years old,

contain organogenous deposits -- graphite gneisses and limestones -- interbedded with sandstones-quartzites. The Madagascar Graphite system dates back 2600 million years. Thus the presence of life is associated with the oldest known stages of formation of the sial envelope, so that this feature is not a reliable criterion for a two-fold differentiation of the Precambrian.

A study of the structure in the Precambrian platforms reveals their many-storied character and the fact that the formation history for Precambrian folding zones is characterized by the same features which mark the post-Cambrian mobile geosynclinal provinces. The early formation stages of such mobile zones are marked by igneous greenstone spilite-keratophyric and porphyry formations and ultrabasics, whereas the terminal stages are marked by various terrigenous flyschlike and molasse formations, and by the same regular stages and series of igneous rocks and associated metallogenies.

There is a tremendous development of useful metal minerals in Precambrian folded systems: 75% of gold and 60% of iron and manganese have been produced from them, along with copper, nickel, cobalt, chromium, uranium, niobium, tantalum, zircon, and rare elements. In the presence of immense granite and migmatite massifs, the metallogeny of these provinces is affected to a great extent, by synclinoria where the sedimentary formations have been preserved.

Zones of early consolidation, with their ancient cores, have been differentiated in the structure of individual Precambrian folded systems. There are blocks representing ancient cores regenerated by younger folding, on a new structural plan. There also are mobile geosynclinal zones with a long development along the same structural plan, wherein several structural stages, of different ages, inherit the same trend, such as in the Buzh (Bug) structural systems of the Ukraine. Finally, there is a radical rebuilding, the superposition of a transverse trend of younger folded systems upon the older.

The initiation of new mobile geosynclinal zones was accompanied by the breaking up of older systems of the lower stage, with the outpouring and building up of igneous formations in the new structural stage. Such sharp changes in the building plan of a new system is characteristic of major stages in the geologic history, as for instance the submeridional superposition of the Saksagan rocks on the northwest trending Bug system. Windows of the ancient lower folding stage are exposed in the Saksagan rock province.

The series of mobile Precambrian zones continue into the Caledonian Paleozoic folding epoch, as shown by the Sayan-Baikal Arch, the Katanga rocks (Katanga geosyncline), and elsewhere. It is commonly difficult to differentiate the Cambrian and the Precambrian, as well as the time of folding in these zones. They embrace the greatest section of the upper transitional interval from the Precambrian to Cambrian.

Nonmetamorphosed uppermost Precambrian beds rest horizontally in the Russian platform, where the consolidation of the folding systems (except for the Timan and Urals) took place in the older epochs.

Thus, there is no basis for the drawing of a special boundary line between the Precambrian and post-Cambrian marking a radical change in the course of geologic phenomena, and for the differentiation of pregeosynclinal and postgeosynclinal post-Cambrian periods, as has been done by some students.

Like the post-Cambrian tectonic cycles, all cycles of mineralization and folding have been characterized by the same regular features, in the last 3.5 billion years of the history of the formation of the earth's sial envelope.

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LENGTHY DEVELOPMENT OF GEOSYNCLINAL FOLDED STRUCTURES IN THE EASTERN PART OF THE CRIMEAN MOUNTAINS[1]

by

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INTRODUCTORY REMARKS

The first proof of a lengthy development of the Donets Basin folded structures was given by N. S. Shatskiy, as early as 1924 [11]. By 1937, he substantiated his concept of a lengthy folding process by more extensive material, in his well-known paper, "Neocatastrophism" [12]. At the same time, V.I. Popov [5, 6], on the basis of data on the geology of central Asia, came to a conclusion on the continuity of tectonic processes. In his 1937 paper, N.S. Shatskiy strongly criticized the concepts of G. Stille and his followers on the alternation of long and quiescent periods of sedimentation with short epochs of intensive tectonic movements known as phases of folding. A substantiated criticism of the Stille canon and of his phases of folding was also offered by V.I. Popov.

Nevertheless, the concept of an alternation of short folding phases and long periods when no folding took place, has gained wide popularity among the Soviet geologists. The prolonged development of folding was accepted by that school of thought as valid only for provinces of domal folding, on platforms and in plunging areas of major folds (Kerch' Peninsula, Apsheron).

V.V. Belousov [1, 2] considers the phenomenon of folding by the provinces of its development, and he divides it into three types: the full, intermediate, and the discontinuous. Only for the latter does he postulate a prolonged development. As for the first two, he states that there are no reasons for their longevity. On the contrary, there is much evidence of their quick appearance and of only a brief phase of development.

According to V. Ye. Khain, all three types of folding are long in developing but not at varied rates. There is a rhythmic acceleration and slowing down in their development processes, with an occasional full stop. This irregularity is different for different types. For linear folds (full folding), periods of slow growth alternate with those of greater intensity in the folding movements, coinciding with breaks in the sedimentation, which V. Ye. Khain calls the phases of accelerated abrupt growth of folding.

In recent years, N.S. Shatskiy [13] again criticized the concepts of Stille, or the neocatastrophists, as he calls the advocates of folding as a manifestation of brief folding phases. He cites evidence of a prolonged folding process in the Neogene of the southeastern tip of the Caucasus. He also notes that folding is an irregular process consisting of long phases, each contributing to the final quantitative change in the general structural plan of that segment of the earth's crust.

One of the authors of this paper [4] has cited evidence of a prolonged process of formation of the Sudak-Karadag folding system of the Crimea, on the basis of material on the Upper Jurassic facies change in the components of these folds.

Nevertheless, this question is still under discussion. The advocates of short folding phases persist in stipulating lengthy folding phases only for platform and domal structures in provinces of discontinuous folding, rather than for the linear intrageosynclinal folds. Specifically, this view is reflected in the last edition of V. V. Belousov's major work [2] on geotectonics. He writes on p. 359: "There are no known well-substantiated instances of changes in the thickness and facies in any definitely full folding which would suggest a slow sedimentary process."

We believe that it is easier to regard a folding process as discontinuous by the nature of the evidence. The demonstration of a prolonged folding process calls for a considerably larger amount of more comprehensive data and for a more careful study of the stratigraphy, lithology, facies, and tectonics. The accumulation of such data is impossible

[1] Dlitel'noye razvitiye geosinklinal'nykh skladchatykh struktur vostochnoy chasti gornogo kryma.

in many places having ancient Paleozoic or Precambrian folding. All this has led, and is leading now, to the negation of the concept of the length of the folding process or else to a doubt in the universality of this phenomenon.

As a matter of face, a lengthy folding process is paramount in importance in the formation of folded structures. The "phases" of folding, as expressed in unconformities, are related not to the folding but to the rising and sinking of large blocks of the earth's crust. The resulting sharp angular unconformities give the impression of a rapid folding process. It should be noted that the concept of a lengthy development in no way precludes the discontinuity and abruptness of the formation process. We believe, however, that the manifestations of such discontinuity have nothing in common with the phases of folding.

In 1954-1955, we studied, in collaboration with G.I. Nemkov, I.V. Arkhipov, M.V. Mikhaylova, and Ye.A. Uspenskaya, the stratigraphy, facies, and folded structures of the eastern part of mountainous Crimea, in connection with the preparation of a geologic map. The detailed data thus obtained on Upper Jurassic facies changes, closely related to the folded structures, have substantiated the above-mentioned conclusions of M.V. Muratov [4] on the length of the folding process. These conclusions can be refined and presented in detail in the light of the new data. The facies changes of Upper Jurassic deposits here, noted by K.K. Focht as early as 1897, were so intensive as to suggest an exceedingly interesting sedimentation picture of that epoch, on the background of the developing linear-type folded structures, i.e., under the conditions of V.V. Belousov's full folding.

This is the reason for our careful study of the data, as the basis of these conclusions.

MAIN FEATURES OF THE GEOLOGIC STRUCTURE OF THE EAST CRIMEAN MOUNTAINS

An early idea of the folded structures of the east Crimean Mountains was presented by V.D. Sokolov [9]. Subsequently, the tectonics of this region was studied by D.V. Sokolov and M.V. Muratov. D.V. Sokolov [8] was first to notice the considerably more complicated development of the Crimean tectonics, than had been supposed before. According to him, the Crimean Mountains are a result of two genetically and structurally different movements: a Jurassic folding and a Tertiary thrust.

D.V. Sokolov believed that Jurassic sediments were folded in two phases: the Kimmeridgian (in the beginning of the Middle Jurassic) and the Andean (early Tithonian). In the Tertiary, the earlier folds were complicated by nappes, with major horizontal rock displacements. As a result, the concept was formed of a rootless occurrence of the Karadag limestone and igneous massifs, as tectonic remnants outside the normal section peculair to the locality of their occurrence. These ideas of D.V. Sokolov were developed especially by N.A. Preobrazhenskiy [7] who believed in the existence of true "charriages" in the structure of this region.

M.V. Muratov [4] has here differentiated several large anticlinal and synclinal folds which he regarded as the eastern plunge of the Tuak anticlinal uplift. He also noticed the younger, probably Tertiary, normal cross faults.

The thick Jurassic section of the Sudak-Karadag region has been gathered into a system of assorted sizes, trending almost latitudinally and complicated by fairly numerous longitudinal and transverse faults. Three principal tectonic structures of the first order have been identified here, with the Tuak anticlinorium in the central position, the Sudak synclinorium to the south and the east Crimean synclinorium to the north (Fig. 1). These three tectonic units make up the core of the Crimean mega-anticlinorium, in its eastern part.

The Tuak anticlinorium is an anticlinal uplift, about 70 km long, trending from Alushta to Planerskoye Village. In the western part, not considered here, the anticlinal axis is uplifted and the apex is deeply eroded. The Taurian series and Middle Jurassic rocks are exposed in the area of Alushta, Rybach'ye and Morskoye. These rocks form a latitudinal system of complex folds, overturned to the south; they make up the core of the anticlinorium.

In the eastern part (east of the Voron River valley) the axial part of the anticlinorium includes, besides the Taurian series and Lower Jurassic, the Oxfordian stage deposits, including the Lusitanian substage, resting with a sharp unconformity on the core of the anticlinorium in a system of folds superimposed on the latter. In the area of the Karadag Mountain group and of Planerskoye, the core plunges sharply, and the anticlinorium closes.

The Sudak synclinorium has been destroyed by the sea to a considerable extent, with only a small segment of it preserved along the shore, on the Meganom Peninsula and in the Sudak area. A number of latitudinally trending folded structures are present within the

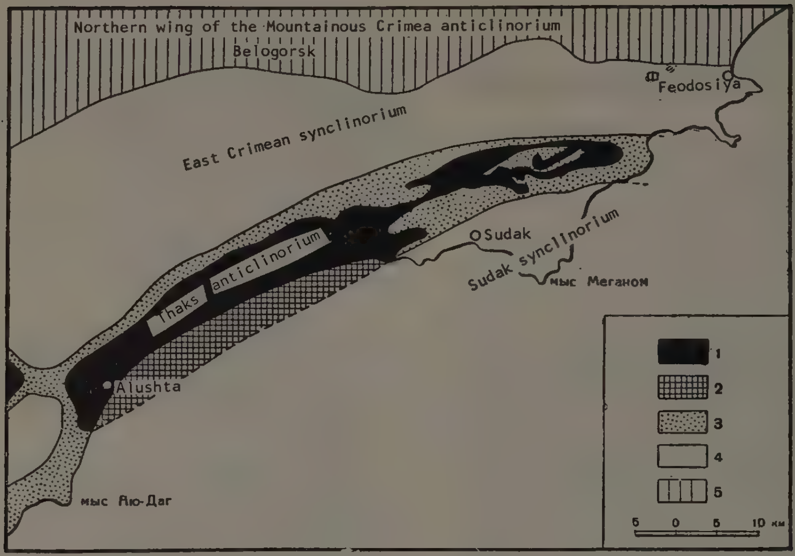

FIGURE 1. Diagram of the distribution of main structural elements in southeastern Crimea.

1 -- The anticlinoria cores, made up of the Taurian series and Middle Jurassic; 2 -- ditto, presumably below sea level; 3 -- limbs and sunken segments of the same anticlinoria; 4 -- synclinoria; 5 -- limbs of the Crimean Mountain mega-anticlinorium.

the synclinorium. The northernmost is separated in the east from the Tuak anticlinorium by a major Echkidag thrust, along which the anticlinorium structures have been somewhat pushed upon the adjacent syncline (Fig. 10).

The east Crimean synclinorium is made up of younger rocks: Kimmeridgian-Tithonian and Lower Cretaceous, with Upper Cretaceous and Paleogene in the eastern part. This latitudinally trending synclinorium occupies a vast area from Simferopol' to Feodosiya. Without describing all of it, we will confine our study to a segment of its southern limb, without which the history of the development of folded structures in the Tuak anticlinorium and Sudak synclinorium would be incomprehensible.

An extremely important feature of the geologic structure of this region is the clean-cut difference in the stratigraphic section of the three main east Crimean structural elements.

A continuous sequence of deposits, from the Bathonian to Tithonian, is involved in the structure of the Sudak synclinorium. The older beds are not exposed here, and not much can be said about them.

Bathonian deposits are exposed in the core of the Kopsel'sk anticline where they consist of argillaceous rocks with siderite concretions and thin sandstones. Upper Jurassic deposits are represented by a very thick, chiefly argillaceous series, divisible into a number of formations. The Callovian stage is represented by shales interbedded with sandstones, limestones, and marls, a total thickness of more than 600 m. They contain the well-known Kopsel'sk fauna of Callovian ammonites. M.V. Muratov [3] also described here such an important middle Callovian index form as Reineckia anceps Rein. However, Callovian deposits change rapidly, going northward. In the limbs of the Perchem anticline and Echkidag syncline (Fig. 2) (separating the Sudak anticlinorium and Tuak anticlinorium), the Callovian rests with a sharp unconformity upon the Middle Jurassic and is represented by yellow, friable sandstones, sandy limestones and marls of obviously shallow facies. The Callovian is missing in the Tuak anticlinorium farther north, which emphasizes the difference in the history of the Tuak anticlinorium and Sudak synclinorium, which became obvious as early as the onset of the Upper Jurassic.

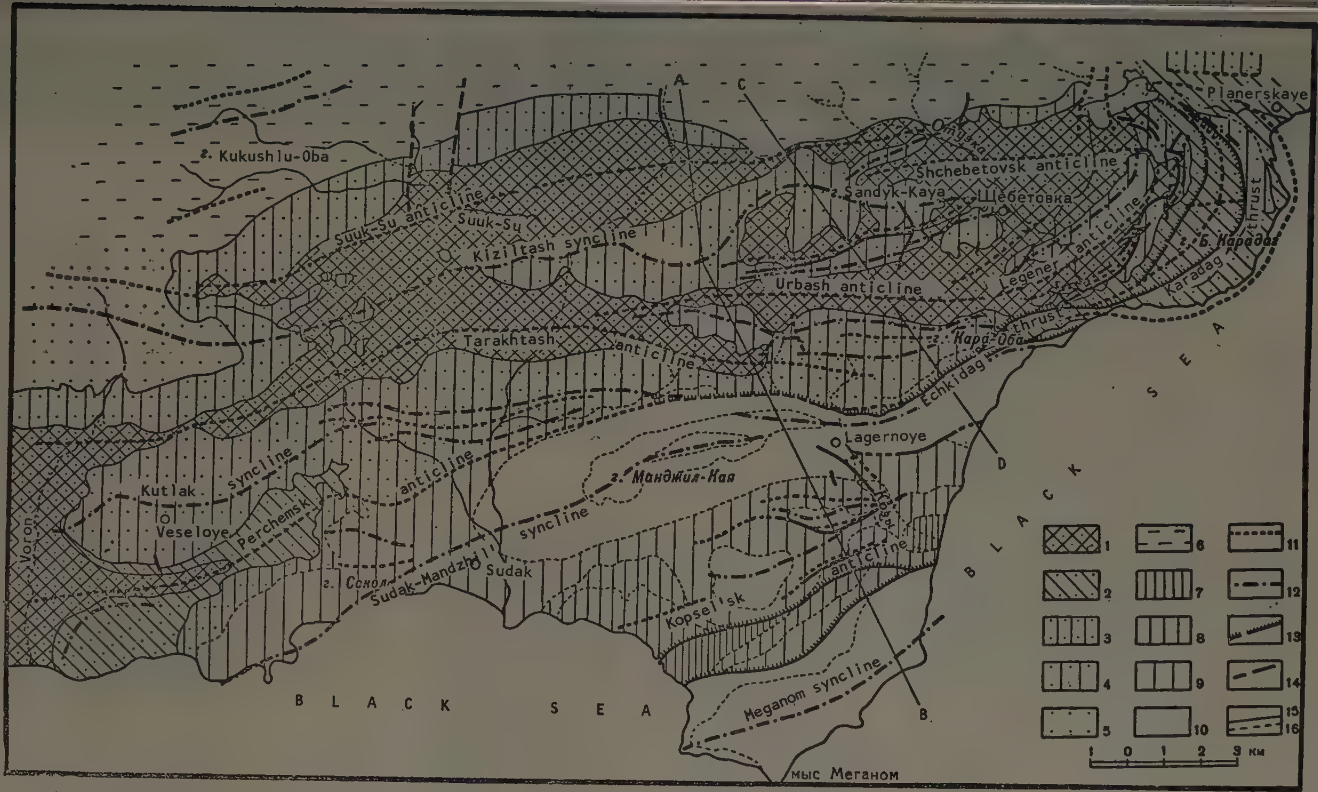

FIGURE 2. Tectonic map of the southeastern part of the Crimean Mountains

The core and apex of the Tuak anticlinorium. Lower structural stage: 1 -- Flysch of the Taurian series in cores of anticlinal uplifts; 2 -- Middle Jurassic deposits. Upper structural stage: 3 -- Callovian deposits; 4 -- Oxfordian and Lusitanian deposits in cores and limbs of synclines; 5 -- Tithonian deposits. Northern limit of the Tuak anticlinorium: 6 -- Tithonian and Cretaceous deposits. The Sudak synclinorium: 7 -- Bathonian mudstones in anticlinal cores; 8 -- Callovian deposits; 9 -- Oxfordian-Lusitanian deposits; 10 -- Kimmeridgian and Tithonian deposits; 11 -- anticlinal axes; 12 -- synclinal axes; 13 -- major thrusts; 14 -- other faults; 15 -- boundaries of unconformities; 16 -- boundaries of conformable occurrence and marker beds.

Oxfordian-Lusitanian deposits are represented by the Karaman formation of shales no less than 800 m thick, with minor intercalations of sandstones, conglomerates, and limestone lenses and with minor massive reef limestones. A special facies of these deposits (the "Sudak formation") has been separated in the Sudak area, on the northern limb of the syncline; it is characterized by the growing importance of siltstones, sandstones, and reef limestones, going toward the edge of the Tuak anticlinorium. The reef massifs in Oxfordian-Lusitanian deposits have their best development along the boundary with the west limb of the Perchem anticline, west of Sudak, where they make up entire mountains and promontories, in the Novyy Svet kolkhoz (collective farm) area. These reef massifs are responsible for the complex relief and the exceedingly picturesque scenery of that area.

Kimmeridgian formations, which change without break to the Lusitanian, consist of shales with layers of siderites and less common sandstones (the Kozy formation), with overlying shales interbedded with sandstones forming a thick flyschlike sequence (the Tukluk formation), total thickness 1000 to 2000 m. An obviously shallow-water facies appears in these deposits along the northern limb of the synclinorium, in the Karshters Mountain area. This facies carries beds of limestones and limestone conglomerates representing the products of disintegration of Lusitanian limestone massifs in the Tuak anticlinorium area (Fig. 8). Conformably resting on the Tukluk formation are conglomerates tentatively assigned to the Tithonian. They are developed in the northern part of the Sudak-Mandzhil' syncline, and they change upward and along the dip to an argillaceous flysch sequence (shales with sandstone layers) which make up the very core of the syncline (Fig. 3). The thickness of the remaining Tithonian sediments is about 300 m. Thus the overall thickness of Upper Jurassic deposits in the Sudak synclinorium reaches 3000 m.

The picture is quite different in the Tuak anticline. Its core is made up of older rocks belonging to the Taurian series and Middle Jurassic. The latter are distributed chiefly along the periphery of the uplift (Fig. 4) and consist of an arenaceous argillaceous flyschlike sequence and black argillaceous rocks of a considerable thickness. On the eastern plunge of the anticlinorium, a thick extrusive sequence (as much as 600 m) appears in the Middle Jurassic. It consists of spilites, keratophyres, tuffs, tuff breccias, etc., of the Karadag formation which is responsible for one of the most beautiful places in the Crimea -- the Karadag Mountain country. It is of interest that obviously littoral sandstones are present among Middle Jurassic rocks along the southern rim of the Tuak uplift, where they form thick lenses [5], most likely the remains of submarine deltas.

Upper Jurassic deposits of the Tuak anticlinorium are quite distinct. They are represented chiefly by Oxfordian-Lusitanian beds, with the Callovian present only along the periphery of the uplift (Fig. 6). As already noted, Callovian deposits here rest unconformably on the underlying rocks and are represented by littoral facies. We have established the presence of extrusives -- Callovian andesites and tuffs -- unconformable upon eroded Middle Jurassic extrusives and closely related to Callovian shallow deposits.

Oxfordian-Lusitanian deposits in the Tuak uplift area are marked by a considerable thickness, as much as 600 to 800 m, and by

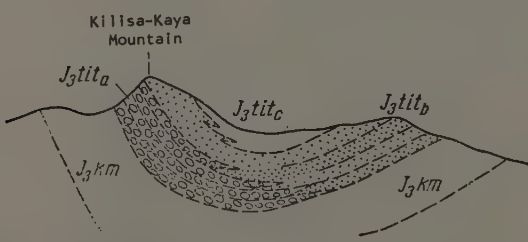

FIGURE 3. Replacement of Tithonian conglomerates by flysch, in the axial part of the Sudak-Mandzhil' syncline.

J_3tit_c -- argillaceous flysch; J_3tit_b -- arenaceous flysch; J_3tit_a -- conglomerates; J_3km -- Tukluk formation (Kimmeridgian).

FIGURE 4. Distribution of Middle Jurassic facies.

1 -- Province of probable slight accumulation of argillaceous deposits with siderites, eroded by the onset of Upper Jurassic transgression; 2 -- shales interbedded with marls and sandstones; 3 -- sandstones; 4 -- extrusives of the Karadag tuffaceous lava series.

the extremely rapid change in facies. These are conglomerates, calcareous sandstones, bedded and massive (reef) limestones, shales with siderite concretions -- all rapidly replacing each other (Fig. 7). We shall take up in more detail the exceedingly complex picture of the distribution of these facies, after a description of the folded structure of the anticlinorium.

Hardly any younger deposits have been preserved in the apex area of the Tuak uplift. Judging from the composition of pebbles in Kimmeridgian-Tithonian conglomerates, a vigorous erosion of the uplift was going on, south and north of the anticlinorium, at the close of the Upper Jurassic. North of Shchebetovka Village, we discovered a transgression of thin Tithonian and Valanginian deposits

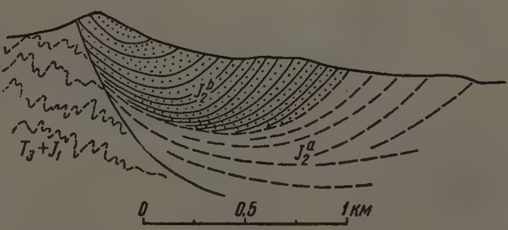

FIGURE 5. Structure of a lense of Middle Jurassic sandstones in the Panaya-Kaya mountain, west of Sudak.

T^3+J_1 -- flysch of the Tarian series; $J_a{}^2$ -- arenaceous argillaceous sequence; $J_2{}^b$ -- Middle Jurassic sandstones.

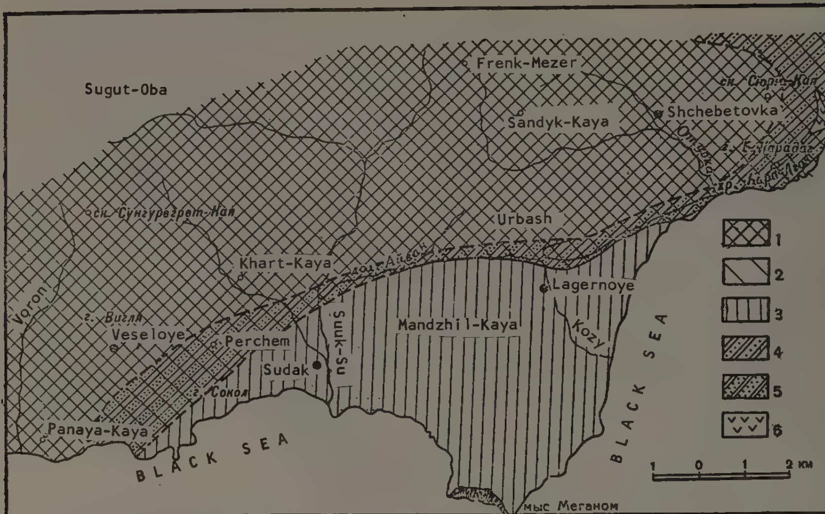

FIGURE 6. Distribution of Callovian facies.

1 -- erosion area, the axial part of the Tuak uplift; 2 -- area of plunge of the rim of the Tuak uplift, with the Callovian unconformable on Taurian and Middle Jurassic deposits (background); 3 -- the Sudak depression area, with the Callovian conformable on the Bathonian. Shales with sandstones and marls; lithofacies along the anticlinorium edges (2): 4 -- shales with marls and limestones; 5 -- sandstones and shales; 6 -- extrusives.

on deeply eroded Lusitanian folds.

The east Crimean synclinorium is made up of rocks ranging from Kimmeridgian and Tithonian to Paleogene. The Kimmeridgian-Tithonian opens with conglomerates (Fig. 9) in a sharp unconformity upon the Taurian series and Oxfordian-Lusitanian along the northern rim of the Tuak anticlinorium. High up in the section, there is a thick (as much as 1500 m) argillaceous flysch sequence of alternating argillaceous rocks, sandstones, and limestones; they change without a break to thick Lower Cretaceous deposits, still higher up. In this segment of the southern limb of the east Crimean synclinorium, Kimmeridgian-Tithonian conglomerates lie with a steep to nearly vertical dip, which becomes gentler toward the north. They are broken up by normal cross faults.

This section, as described from the principal structures in southeastern Crimea, clearly demonstrates the close relationship between its nature and the major structural forms (Figs. 6 to 9). An explanation of this phenomenon is to be sought exclusively in the character of development of these structural forms. Their lengthy development and slow growth are not to be doubted. In addition, these major structures -- the Tuak anticlinorium and the Sudak and east Crimean synclinoria -- definitely belong to linear geosynclinal folding of the full type. This is readily seen from their structure as correlated with V.V. Belousov's illustrations of that type of folding [2].

There are many such instances of similar major long-developed co-sedimentary structures in geosynclinal provinces. They have been described from central Asia, by V.I. Popov; from the Urals, by A.V. Peyve; from the Caucasus, by V.Ye. Khain and D.S. Kizeval'ter, etc.

A special interest of the area under study lies in the fact that individual linear folds of the Tuak anticlinorium, undoubtedly of a full folding type, also display a very close relationship between the distribution of facies -- which we have studied in detail -- and the individual folded forms; in other words, they are slow growing co-sedimentary structures. For a proof of this, we shall consider in more detail the tectonic structure of main zones of folding.

FIGURE 7. Distribution of Oxfordian and Lusitanian facies.

1 -- erosional areas in the apex of the Tuak uplift -- cores of young, slowly-growing anticlinal folds, with Taurian argillaceous flysch exposed and eroded in the process of the anticlines' growth; 2 -- plunging areas in the apex of the Tuak uplift, with Oxfordian and Lusitanian deposits unconformable on older rocks; 3 -- same, with these deposits conformable on the Callovian; 4 -- the supposed east Crimean trough with argillaceous facies; 5 -- the Sudak trough -- shales with sandstones, conglomerates, and coral limestones. Present exposures of lithofacies: 6 -- reef limestones; 7 -- conglomerates; 8 -- limestones; 9 -- shales.

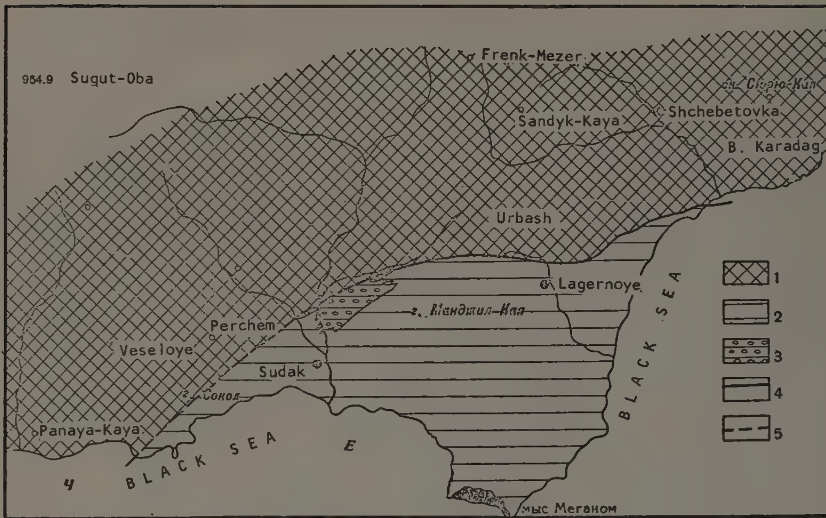

FIGURE 8. Distribution of Kimmeridgian facies.

1 -- erosion area, the axial part of the Tuak uplift; 2 -- the Sudak trough; flysch, with Kimmeridgian deposits conformable on the Oxford-Lusitanian; 3 -- the Karshters facies: calcareous-conglomerates, limestones; 4 -- fault boundaries of facies zones; 5 -- boundaries of facies zones.

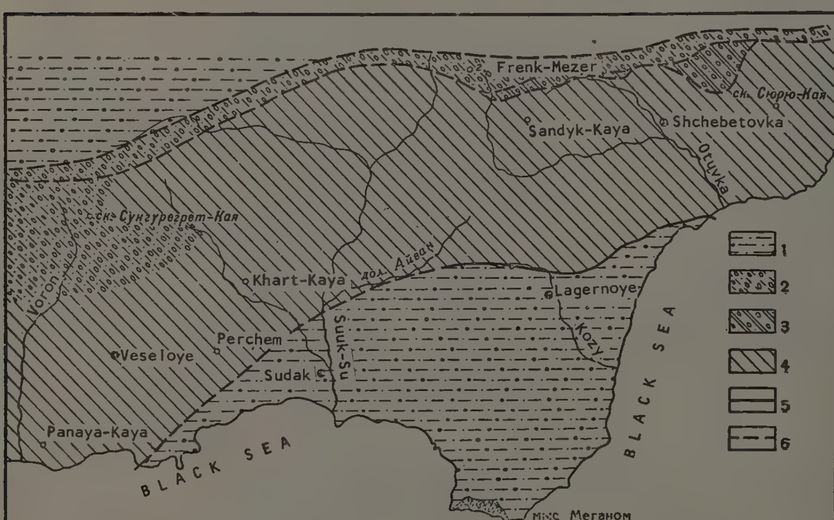

FIGURE 9. Distribution of Kimmeridgian-Tithonian facies.

1 -- The Sudak and east Crimean troughs: argillaceous flysch, conglomerates, and sandstones in the Sudak trough, conformable on the Oxfordian-Lusitanian; 2 -- conglomerates and sandstones, unconformable on the Oxfordian-Lusitanian; 3 -- same, unconformable on the Taurian series; 4 -- area of the Tuak uplift, at times involved in the sinking; 5 -- fault boundaries of facies zones; 6 -- boundaries of facies zones and lithofacies.

THE STRUCTURE OF THE TUAK ANTICLINORIUM

Two structural stages, separated by a pre-Callovian unconformity, are well defined in the apex part of the Tuak anticlinorium.

The lower structural stage (Fig. 10) which includes the Taurian series and Middle Jurassic rocks, is marked by chiefly argillaceous flysch facies forming extremely complex small folds which almost conceal the larger structures. Nevertheless, it is possible to establish from the change in individual members and from the continuity of certain beds that, although the prevailing strike of the lower structural stage beds generally coincides with the trend of Upper Jurassic folds, the position of Upper Jurassic rocks on most diversified layers of the lower structural stage suggests a sharp unconformity between them. It is very significant that, as early as the Middle Jurassic, the edge part of the Tuak anticlinorium was an area of the accumulation of shallow littoral arenaceous deposits, along its southern rim (Fig. 4).

The upper structural stage rocks form a system of linear folds, trending nearly latitudinally on the whole, locally intensively faulted. All of this structure is superimposed on an older Early to Middle Jurassic folding.

The upper structural-stage folds are of a linear type, with strongly undulating axes, which renders them like structures transitional to brachifolds. The larger folds are complicated by domal and cuplike forms and by linear folds of the second and lower orders. The upper-stage folds are asymmetrical, with the axial planes tilted north, in the northern part; and south in the southern part. Thus the anticlinorium structure has acquired a fan-shaped aspect (Fig. 11). The southerly overturning is especially well expressed along the southern rim of the core, in the Echkidag Mountains to the east, where the large Echkidag thrust marks the southern boundary of the Tuak uplift (Fig. 10). The structure of the Tuak anticlinorium apex is very complex. Three main anticlinal zones can be differentiated here: the Suuk-Su, Tarakhtash, and Perchem-Karadag. In the area of Karadag and Planerskoye Village, the Perchem-Karadag folded zone veers sharply northeast and north, as if including the two adjacent zones of the north (Fig. 2). Each zone is as wide as 3 km and more than 35 km long. Each of these main zones consists of several large anticlines and synclines which we assign to the first-order folds. Some of them are complicated by longitudinal and transverse folds of the second order and by minor folds. A characteristic feature of this folding system is en echelon arrangement of the first order folds, crossing the main structures at a sharp angle.

The structure of the two northern zones is simpler. The Suuk-Su anticline is asymmetric along most of its trend, with a very steep, usually overturned, northern limb. The Kizyltash syncline is more symmetric, with fairly steep, although low, limbs and a broad, gently wavy bottom. Although generally linear, this fold is differentiated into individual brachimorphic structures with variously oriented axes of the second-order folds. The absence of the outward overturning of peripheral folds, as a general phenomenon, is characteristic. The axial planes are tilted either north or south.

Much more complicated is the structure of the southern folded zones. Here, the folds are interlaced so closely that only a tentative differentiation is possible. The folds are most compressed in the east where the combined width of all three zones is not more than 3.5 km. They spread out in the west, to 6 to 7 km. The axes of the folds rise to the west and east, bringing ancient rocks to the surface. The Tarakhtash anticline is the most elevated; it forms the crest of the Tuak uplift, as if emphasizing its asymmetry. The Perchem-Karadag folds are very complex, which is probably the result of their peripheral position, adjacent to a very deep synclinorium. The southernmost peripheral folds are clearly asymmetric and definitely overturned to the south, toward the Sudak synclinorium. Major thrust faults are developed in the eastern part of the Tuak uplift, along its southern rim and the eastern closure. The largest of them, the Echkidag thrust, branches in the east into a system of the Karadag thrusts, with the formation of large tectonic "scales." The folding here, too, becomes much more complicated. Lusitanian limestones, in a long stretch, stand on edge or else are overturned. Instead of solid limestone massifs to the west, only isolated, comparatively small reefs occur here. It is they that form most of the "rootless" limestone massifs of the Karadag area.

According to our data, these massifs are not "rootless." The area of their distribution is marked by the increasing significance of argillaceous members, numerically predominant in the Upper Jurassic section, and by a very sudden and sharp change in facies. Present in the Upper Jurassic here are compressed folds, not at all typical of this interval in the west. The folds have been complicated by numerous faults -- a result of the breaking-up of competent beds involved in the folding of argillaceous beds. It is the wedges of these competent limestone beds that present the so-called "rootless" massifs of the Karadag.

Middle Jurassic extrusives of the Karadag tuffaceous lavas are exposed in the Karadag.

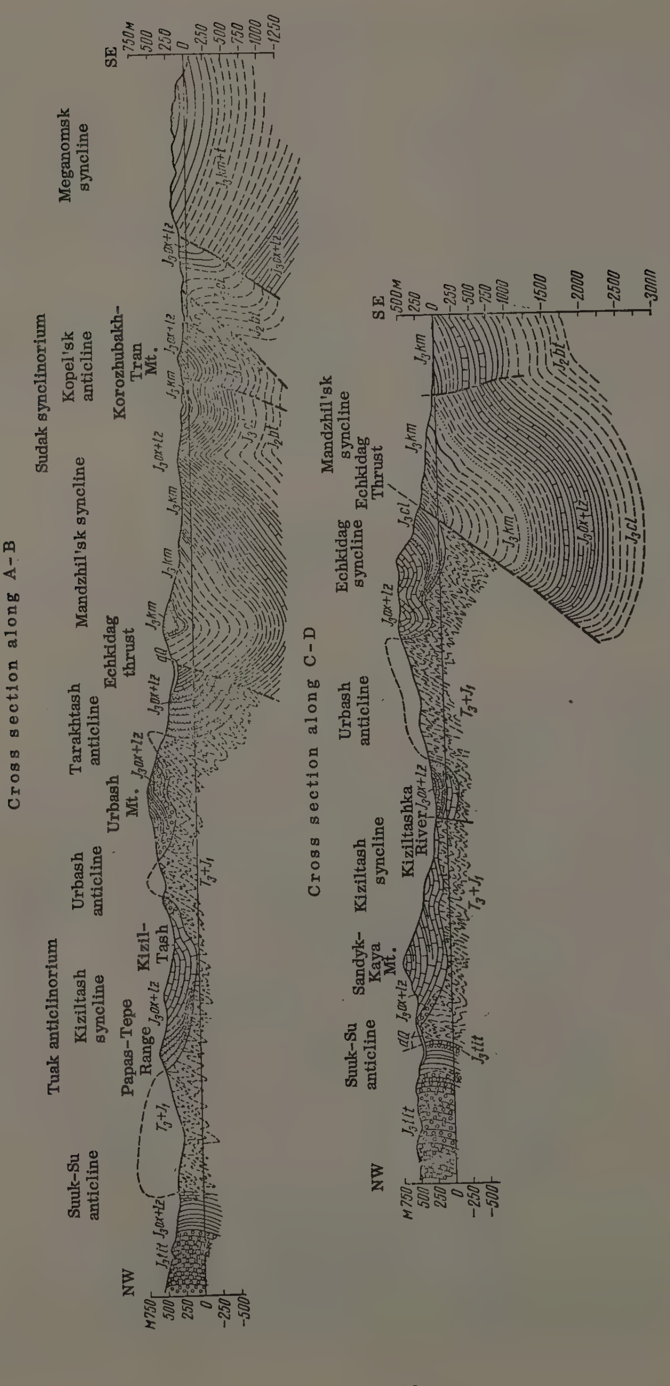

FIGURE 10. Cross sections of the Tuak anticlinorium and Sudak synclinorium.

$T_3 + J_1$ -- Taurian series; J_2bt -- Bathonian stage; J_3cl -- Callovian stage; $J_3ox + lz$ -- Oxfordian and Lusitanian stages; J_3km -- Kimmeridgian; J_3tit -- Tithonian; $J_3km + t$ -- Kimmeridgian and Tithonian.

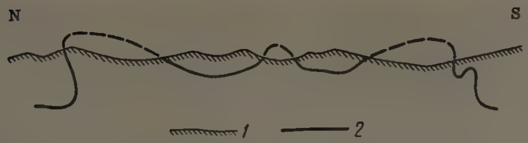

FIGURE 11. Structural diagram of the Tuak anticlinorium.

1 -- Surface; 2 -- level of Upper Jurassic deposits.

They have been pushed up, on a steep thrust, over Callovian beds which rest with a sharp unconformity on an eroded surface of Middle Jurassic structures.

THE STRUCTURE OF THE SUDAK SYNCLINORIUM

Unlike the Tuak anticlinorium, the nonsubmerged, very small segment of the Sudak synclinorium is made up of a thick and complete sequence of conformable rocks, from the Bathonian to the Tithonian. Two synclinal and one anticlinal zones are clearly distinguishable in this synclinorium. In the north, there lies the Sudak-Mandzhil' syncline which is a comparatively simple asymmetric deep fold whose northern limb is broken up by the Echkidag thrust. South of it, there lies the Kopsel'sk anticline, complicated in its apex zone by domal uplifts separated by depressions. The southernmost Kopsel'sk anticline is more complex in its structure, being made up of a number of minor and steep folds which are broken up by thrusts associated with the relative southward ovement. Correspondingly, the folds are overturned to the south, toward the Black Sea (Fig. 10).

THE DEVELOPMENT HISTORY OF THE FOLDED STRUCTURES

A study of the facies changes in Oxfordian-Lusitanian deposits in the Tuak anticlinorium area (Fig. 7) yields very important data on the progress of folding in Upper Jurassic beds. With all the complexity of the facies' distribution here, their close relationship with the arrangement of the above-described folds in the anticlinorium apex is especially convincing. The zonal distribution of definite facies groups can be traced as regularly outlining the folds.

Narrow belts of conglomerates, locally replaced by sandstones, stretch along the periphery of synclinal troughs filled with Oxfordian-Lusitanian deposits, at their boundary with anticlinal windows of the basement, made up of the Taurian series and the Lower Jurassic. A reef zone (mentioned before) is usually located farther away from the depression rim, very seldom near it. These reef zones are especially well expressed on the northern limb of the Sudak syncline, in the interior of the Kizyltash syncline, and on the southern limb of the Tarakhtash and Perchem anticlines. Large reefs are also present along the periphery of the Shchebetovsk and Legener anticlines and along the southern rim of the Urbash

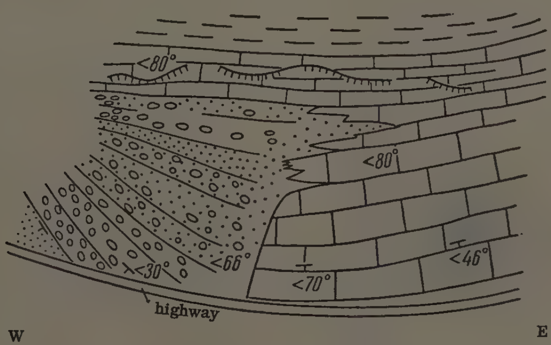

FIGURE 12. The character of replacement of Oxfordian limestones by conglomerates, near Shchebetovka Village (on the highway to Sudak). Aerial view.

anticline. The very abrupt change of reef limestones to coarse clastics, suggesting the contemporaneous formation of both, is very well expressed. There are instances of juxtaposition of clastic accumulations and previously formed steep to sheer reef slopes (Fig. 12). This bears testimony to a complex and sharp bottom relief and to the filling of individual troughs with coarse clastic material. Finally, the interior part of synclinal troughs is filled with argillaceous facies ("Ayvan formation") usually consisting of very fine, well sorted, homogeneous, unstratified sediments apparently accumulated in small interior lagoons of very quiet waters.

Of all these facies, we are most interested in conglomerates. The influx of coarse clastic material from the north, east, and south during Oxfordian time, is ruled out, because argillaceous sediments were being deposited on all sides of the Tuak uplift at that period. A possible source of sediments is the more elevated western part of the uplift. However, a more detailed study of the facies' distribution in the area does not support this theory. Conglomerates occur in the west (Kutlak syncline) as sporadically as they do elsewhere. They are developed in the east (Shchebetovka and Legener folds) to the same extent as in the west, with the clastic material not any coarser or less rounded, and without any other evidence of the proximity of a western source. Nor do the conglomerates form an unbroken area of deposition or anything suggesting an alluvial fan. On the contrary, they definitely fringe the anticlinal ridges of the ancient basement in sinuous bands, with reef limestones constantly accompanying the conglomerates. Finally, considering the particularly monotonous composition of the local pebbles consisting almost exclusively of Taurian sandstones, abundant in the argillaceous section which makes up the anticlinal cores, the only conclusion is that the latter -- in the most elevated parts of the largest folds of the Tuak anticlinorium -- were the source of the clastics. Considering further the great thickness of the conglomerates (as much as 450 m) and reefs (as much as 800 m), it appears that these anticlines supplied the clastic material during the entire Oxfordian-Lusitanian deposition. This is possible only on the assumption of a very slow and long growth of these anticlinal and synclinal folds, commensurate with the sedimentation rate. The complex relation of facies assemblages throughout the section, with an alternate expansion and contraction of the deposition area, suggest a discontinuous and abrupt growth of the folds. As a result, the substantial length of the process and the co-sedimentary character of the typical linear secondary folding within the Tuak anticlinorium in the Oxfordian-Lusitanian, are quite obvious.

There are data on a lengthy development of other structures, as well, especially in the transitional zone of the southern limb of the Tuak uplift.

The sharp difference in the Middle Jurassic and Callovian section, on either side of the Karadag-Echkidag thrusts, suggests their pre-Callovian origin. It is also obvious that their development went on, probably intensified, throughout the Upper Jurassic.

It is particularly interesting that there is an abrupt thinning of the Callovian section in the peripheral south Echkidag anticline[2] of the Tuak anticlinorium, whose apex is made up of Callovian deposits. This thinning is accompanied by the disappearance of argillaceous intercalations, so that the section consists of conglomerates and sandstones, whereas argillaceous deposits predominate in the adjacent synclines, both to the north and the south. This extremely important fact suggests a contemporaneous formation of the anticline and the accumulation of Upper Jurassic deposits. An indirect confirmation of that is found in the Perchem anticline where there is a considerable complication in the folding of Callovian beds, and an appreciable angular unconformity with the overlying Oxfordian. It is clear that both the development of the Tuak anticlinorium and the growth of its individual folds begin as early as the Callovian.

A study of the northern limb of the anticlinorium shows that the further growth of the folds continued until the deposition of Tithonian conglomerates. This limb is made up of steep to overturned Lusitanian and Tithonian beds, with the Tithonian nonconformable on the Lusitanian, and in many places directly on the eroded cores of the northernmost peripheral anticlines. The unconformable position of the Cretaceous and Eocene suggests that the anticlinorium kept growing in the Tithonian and later, although with a lower intensity.

Let us now consider the overall course of the development history for the entire structure of this Crimean area.

At the onset of the Alpine cycle -- in the Triassic and Lower Jurassic -- all of the Crimean Mountains presented a segment of a vast flysch geosyncline. There is no evidence of the proximity of the edge of that trough, in our area. At the close of the Lower Jurassic and the beginning of the Middle Jurassic, southeastern Crimea underwent a major uplift

[2] 1.7 km northeast of Lagernoye Village (see Figs. 2 and 6).

accompanied by a deep erosion of the Taurian series. The sharply unconformable position of Middle Jurassic deposits on various folded structures of the Taurian series points to a pre-Middle Jurassic folding. The first sign of the Tuak uplift appears in the Middle Jurassic, with an incipient differentiation of the sedimentary basin into two parts: the northern, in the Tuak uplift area where comparatively thin sandstones and argillaceous and arenaceous deposits were probably accumulated; and the southern, along the periphery of the uplift, with chiefly argillaceous sediments. In the area of the eastern plunge of the anticlinorium, at the Karadag, the Karadag volcanic complex was formed, probably associated with the cross faults, east of the uplift.

At the close of the Middle Jurassic, the intensive warping of the entire province gave place to an uplift in the Tuak geanticline area, and to a retreat of the sea. From then on, the geanticline shape is definite, outlined by the unconformable position of the Upper Jurassic beds. Marine conditions persisted in the Sudak geosyncline. Here, the deposition of a thick argillaceous section continued throughout the Bathonian, and without a break into the Callovian.

At the beginning of the Upper Jurassic, the eastern and southern rims of the Tuak geanticline were again involved in a downwarping. The southeastern part of the uplift, along with its eastern part in the Karadag area, was submerged at the beginning of the Callovian. The sinking of the southern limb apparently proceeded along fault planes, so abrupt is the facies change from argillaceous beds of the downwarp to sandy conglomerates which fringe the uplift. In the eastern part, the sinking was accompanied by new volcanic flows. In the Oxfordian, the downwarp embraced all of the remaining eastern part of the Tuak uplift. The process, however, was not that of a gradual and uniform sinking of the entire area.

The above-cited data on the structure of the anticlinorium folds demonstrate that the general sinking of the Tuak geanticline was accompaned by the growth of individual minor uplifts within it. They were the nuclei of future anticlinal folds.

It should be emphasized that despite the general sinking of the entire area and the establishment of marine conditions, the difference between the Tuak and Sudak zones persisted throughout the Oxfordian-Lusitanian. Shallow and littoral facies predominate in the Tuak zone, with an extremely motley and varied picture of physical geographic conditions. The abundance of pebbles of exclusively indigenous rocks is related to the presence of islands made up of Taurian rocks and representing the cores of slowly developing anticlinal folds. Thick deposits of pebbles and sand, change extremely rapidly to limestone reefs of assorted sizes, which fringed large lagoons gradually filled with argillaceous material.

The more uniform environment of an open sea with the accumulation of argillaceous sediments prevailed in the Sudak geosyncline, south of the Tuak uplift. Even here, however, the frequent change in facies throughout the section, from coral reefs to pebble beds, suggests great mobility of the sea bottom and the formation of local uplifts.

A feature of the Oxfordian-Lusitanian interval is the lack of difference in its thickness between the geanticlinal and synclinal belts. On the contrary, there is local thickening of it along the southern edge of the Tuak zone. It appears that there were special exogenetic physical geographic conditions operative here in bringing about this increase in thickness, such as the growth of the reefs and the lingering of the clastic material in the reef lagoons in the north; and the lack of adequate supply of sedimentary material in the south, in the somewhat deeper and, consequently, more downwarped Sudak downwarp. The latter, therefore, was not fully compensated by sedimentation.

Thus the Tuak geanticline remained relatively higher than the Sudak downwarp, under the Oxfordian-Lusitanian conditions of a general and intensive sinking of the entire system.

The difference between the two zones became even greater in the Kimmeridgian and Tithonian. In the Kimmeridgian, the Sudak geosyncline underwent intensive downwarping with the formation of an argillaceous flysch, as much as 1000 m thick. Taking into account the increasingly deep-water nature of the sediments, it can be stated that the amount of downwarping was considerably more than 1000 m. In the Tuak anticlinorium area, judging from the presence of Kimmeridgian conglomerates of Lusitanian limestones along its southern fringe, there was at least a local intensive erosion. Kimmeridgian sediments were altogether lacking here, at the onset of the Tithonian.

It appears, then, that the Tuak geanticline was rising during the Kimmeridgian. Just as definite is the continuation of this process in the Tithonian, although individual segments of the uplift, along its northern rim, were involved in downwarping, with the accumulation of conglomerates. The unconformable position of the Tithonian directly on ancient cores of northern anticlines is an incontrovertible evidence of Oxfordian-Lusitanian beds being

folded by that time. These folds continued their development in the Tithonian. Especially well expressed is the growth of the Shchebetovka and Koktebel' anticlines. After the Tithonian uplifts, they were once more covered unconformably by Lower Cretaceous sediments. In the meantime, the adjacent geosyncline underwent uninterrupted downwarping with a conformable deposition of all Upper Jurassic stages.

The east Crimean geosyncline continued in its development, throughout the Lower Cretaceous. The Tuak geanticline maintained its elevated position. The unconformable position of the Lower Cretaceous on the northern peripheral structures of the anticlinorium suggests that a steep flexure was formed there, as early as the Tithonian, with its core thrust over the northern limb.

The subsequent structural development of southeastern Crimea cannot be described in detail, because of the lack of younger deposits. In the Upper Cretaceous to Neogene, east Crimea was involved in a general uplift culminating in the formation of the Crimean mega-anticlinorium toward the close of the Tertiary. The subsequent development and final formation of the Sudak-Karadag folded system, as we know it, belongs to that time.

CONCLUSIONS

The detailed material on the facies and the tectonic structures of Upper Jurassic sediments in the eastern part of the Crimean Mountains provides a substantial basis for the following three main conclusions:

1. It has been clearly established that the development of large forms of full linear folding (Tuak anticline, Sudak syncline) was of long duration, continuing from the Bathonian to Tithonian.

2. Also clearly established is the long development of secondary folds in Oxfordian-Lusitanian beds; these folds complicate the structure of major linear folds. The length of their development is commensurable with that of the sedimentary process responsible for the formation of the folded sequences, and is fixed by the distribution of facies and by local unconformitites.

3. The detailed data on Middle and Upper Jurassic folding in east Crimea provide telling evidence for the assertion that linear-type folds are not formed in isolated brief stages, the so-called phases of folding, but rather develop at a slow, although discontinued, pace over a very long period of time, involving entire epochs. As to the "phases," they represent stages of more general major oscillatory movements accompanied by a more or less intensive erosion of earlier structures within the uplifts, with the emergence of an unconformity at the base of transgressive series of the next sedimentary cycle.

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FACIES ENVIRONMENT OF LOWER CARBONIFEROUS COAL-MEASURES ACCUMULATION THE DONETS BASIN[1]

by

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Facies environments of the lower Carboniferous coal-measures accumulation in the Donbas are identified and described on the basis of detailed lithological observations.

The author concludes that the entire industrial coal-bearing capacity is related to a stable development of a lagoonal environment.

Graphs are presented to illustrate the marsh facies content as a function of the development of a lagoonal environment; and the coal-content coefficient as a function of the marsh facies content. The author deems these data useful as a means of prediction.

* * * * *

The basis of this paper is the data of a detailed geologic study by the facies-cyclic method of analysis, carried out in 1954-1957, in the eastern half of the shallow occurrence of lower Carboniferous deposits in the Donets Basin (Fig. 1). The field observations were done in cooperation with K. F. Kurilova, a lithologist of the Artemugleogeologiya Trust, on cuttings and cores from boreholes. A full section of the coal measures was studied in detail in the following four districts (from west to east): Dobropol'ye, Mezhevaya-Volch'ya prospect, south Donbas, and Karavannaya prospect.

In the geologic prospecting practice of the Artemuglegeologiya Trust, coal-bearing formation C_1^3 is subdivided into two members: the lower ($C_1 - C_5$ limestones), assigned to lower Visean; and the upper ($C_5 - D_1$ limestones), assigned by most students to the Namurian. All consistently workable coal beds belong to the lower member. The overall thickness of the coal measures in the Dobropol'ye and Karavannaya-prospect districts reaches 1, 100 m; it decreases sharply going toward south Donbas, where it is about 600 m thick. In a very rough approximation, the coal-bearing capacity is in inverse ratio to the thickness. In the Dopropol'ye and Karavannaya-prospect districts, there are no more than ten coal bèds, with only a few, locally attaining a workable thickness. As against that, there are 45-50 coal beds in the south Donbas district, 12 of which are consistently workable.

The reasons for such abrupt changes in the coal-bearing capacity were not clear, and the problem was given priority as a basic practical study project.

There had been hardly any detailed lithological work in the production area of the Trust. A paper of V. V. Koperina should be mentioned [4], in which she published a comparative description of facies and types of coal accumulation in the Donbas lower and middle Carboniferous.

For the lower Carboniferous, V. V. Koperina only studied a single section in the south Donbas coal measures, which prevented her from describing their lateral changes. V. V. Koperina's data on rocks, facies, and the sedimentation conditions for lower Carboniferous coal measures are of a general character and contradict in many respects our data.

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[1]Fatsial'nyye obstanovki nakopleniya uglenosnoy tolshchi nizhnego karbona Donetskogo basseyna.

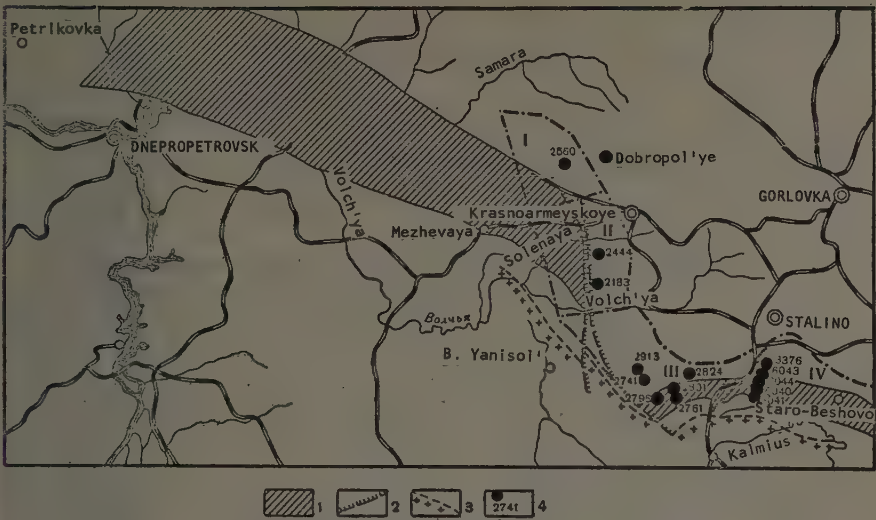

FIGURE 1. Index map of exploratory areas and test holes of the detailed lithologic study of a shallow lower Carboniferous coal-bearing formation C_1^3, in the Donbas.

1 -- area of shallow coal-bearing formation C_1^3; 2 -- major faults; 3 -- crystalline rocks; 4 -- test holes. Numerals: I - Dobropol'ye district; II - Mezhevaya-Volch'ya prospect; III -- South Donbas district; IV -- Karavannaya prospect.

ROCKS AND FACIES

In the area under study, the coal measures are mostly siltstones with subordinate sandstones represented chiefly by very fine to fine-grained varieties. Coarse-grained sandstones and conglomerates are generally very restricted. Just as poorly developed are argillaceous rocks, almost always carrying some silt. The coal content in the lower member locally reaches 3%. Limestones are represented by thin beds accounting for not more than 0.3% of the entire lower-member thickness; their content in the upper member increases as much as 2.5%. The higher limestone content and the lower coal content are the main differences between the two members. The insignificant content of coarsely clastic and fine suspension rocks is a specific feature of the lower Carboniferous coal measures, which sets them apart from the productive middle Carboniferous formations.

The uniform nature of the rocks is enhanced by a comparatively uniform facies composition as expressed in the extremely small local distribution of sandy fluvial facies and in the subordinate character of very fine-grained open-sea sediments, with a normal marine fauna. The best developed are shallow marine facies and those of a sheltered littoral zone with embayments and lagoons, into which river loads were periodically dumped. The latter were reworked by wave action to form accumulations in the form of submarine deltas and, much less commonly, of bars. Of importance are terrigenous deposits of marsh facies in extremely shallow, mostly relict basins showing evidence of stagnation and of being overgrown with autochtonous vegetation. They were formed in periods of maximum regression, over the emerging land relief, along with subordinate lacustrine facies.[2] The relative content of different facies in the lower and upper members, by districts, is given in Table 1.

These data differ substantially from those cited by V. V. Koperina who recognizes only four facies: bars and spits, lagoons and embayments, shallow littoral, and marshes. She assigns deposits of coastal basins to the shallow marine facies. These basins are separated from the sea by bars and spits; i.e., they are in fact lagoonal.

Thus she writes, "Sand bars may rise to the sea level and isolate a belt of near-shore shallow water in the shape of a lagoon oriented along the coast" [4, p. 15].

[2]See "Atlas of Lithogenetic Types of Middle Carboniferous Coal Measures in the Donets Basin" [1] which contains a description of facies and of most genetic rock types also occurring in the lower Carboniferous of the Donbas.

Table 1

Facies composition of coal-bearing formation $C_1^{\ 3}$

Districts	Lower member				Upper member			
	Dobro-pol'skiy	Mezhevaya-Volch'ya prospect	South Donbas	Karavannaya prospect	Dobro-pol'skiy	Mezhevaya-Volch'ya prospect	South Donbas	Karavannaya prospect
Facies	Borehole 2560	Borehole 2183	Borehole 2801	Borehole 6040 + 6041	Borehole 2560	Borehole 2183	Borehole 2824	Borehole 6043 + 6044 + 3376
Open sea, M	8.1	12.9	5.2	6.7	18.0	8.9	17.8	12.1
Shallow sea, MM	23.1	11.3	11.8	20.8	27.5	21.8	29.9	28.6
Embayment, Z	14.1	5.0	0.1	8.0	21.7	22.2	17.9	20.1
Lagoon, Z	20.6	26.3	29.2	32.7	10.5	25.2	15.5	9.2
Bars, B	0.8	3.1	3.0	2.3	1.5	—	4.5	2.3
Submarine delta, D	4.6	3.2	13.0	9.5	7.3	8.2	—	5.5
Fluvial, R	0.1	0.4	—	—	0.5	—	—	3.6
Lacustrine, O	13.8	3.1	7.9	1.1	8.1	4.6	3.2	8.1
Marsh, B	15.0	34.7	29.7	18.9	5.0	9.1	11.2	10.6
Coefficient of Total coal content	0.16	1.69	2.52	0.21	0.42	0.67	0.09	0.02

As it appears from our data, she grossly overestimated the significance of the bar facies and underestimated that of the marine facies. This has led to an erroneous interpretation of the sedimentary environment of the lower Carboniferous coal measures, wherein she sees a complete similarity with the Vorkuta deposit of the Pechora Basin.

A correlation of the lower and upper members by districts shows a higher content of marsh and lagoonal facies in the lower member; and of the marine, shallow-marine, and embayment, in the upper member. It appears as though lagoonal facies are the most favorable for a comparatively wide development of marsh facies and for the formation of coal. However, it is readily seen that the relationship between the lagoonal and marsh facies is not quite firm. Indeed, the lagoonal facies content in the lower member, for the Dobropol'ye and south Donbas districts, is greater than in the upper, by a factor of about two, with a factor of three for the marsh facies. For the Mezhevaya-Volch'ya prospect, the lower member carries four times more marsh facies, with about the same content of lagoonal facies for both members. The ratio changes sharply again, in the Karavannaya-prospect district.

A correlation by the districts shows that the Dobropol'skiy and Karavannaya-prospect districts differ from the central ones by their substantially greater content of shallow marine and embayment facies and by their lower marsh-facies content. Their own facies composition is fairly similar. The relationship between the marsh and lagoonal facies is fully obliterated, apparently because of a narrower range of their areal change as compared to their change with time, in the passage from the lower member to the upper.

A correlation of individual intervals (major cycles) of the lower, properly coal-bearing member, by districts, shows that the marsh facies does not exhibit a consistent relationship, either with any other facies or with their groups. In any individual example, the coal-content coefficient and the content of marsh facies correspond, on the whole, to a specific ratio of the facies; on the other hand, a similar facies ratio is observed for intervals substantially differing in their coal content.

Thus the establishment of sharp facies differences between the members, and of smaller but still definite areal changes, gives no definite clue to the causes of change in the coal content; nor does it promote the

facies analysis as a means to that effect.

Positive results were obtained only after a study of the paragenetic combination of facies within the sedimentary cycles, wherein the role of different sedimentary environments in the coal accumulation has become clear.

THE STRUCTURAL CHARACTER OF SEDIMENTARY CYCLES

The lower Carboniferous coal measures have a well-defined cyclic aspect in their sedimentation. The top of a coal bed carries mudstones or fine-grained siltstones with a marine or slightly brackish fauna. There are a few beds of limestones or calcareous siltstones, usually associated with the base of the terrigenous member. This rock sequence corresponds to the maximum sea transgression and reflects a peripheral -- the farthest away from shore -- segment of the coal-accumulation zone (Fig. 2, subzone 1). Higher up in the section, the rocks get coarser -- to medium-grained siltstone; their faunal content decreases and so does their degree of preservation; the amount of plant remains increases, along with the evidence of intensified wave action. The weight of evidence is in favor of an approaching coast line and of the transition to a zone constantly affected by wave and current action (Fig. 2, subzone II).

The above-named rocks are represented chiefly by coarse-grained siltstones, or else by an alternation of siltstones of different grain size, and by very fine to fine-grained sandstones (Fig. 2, subzone III). This part of the section is marked by a progressive increase in plant remains and in stratification enhanced by ripple marks of both the current and the ebb-and-flow type; by the traces of boring organisms, tracks of worms and other, commonly unidentifiable creatures; and by the evidence of minor modeling of the bottom by currents, and by vegetation brought in from the adjacent land. There also is evidence of drying up, rain drop marks, etc. All of this points to the accumulation of sediments in the immediate proximity of the coast, with an active participation of offshore waves and currents. The section terminates with siltstones carrying abundant plant remains, including roots, the "curly" fossil soil, and the coal itself (Fig. 2, subzone IV). This sedimentary sequence corresponds to the landward, predominantly marshy zone of coal-bearing deposits.

This scheme presents in a most general way the sequence of deposition of different coal-bearing subzones of a sedimentary cycle. It is locally complicated by the wedging-in of thick facies of river sands, more or less reworked by waves and currents, in nearshore basins, and locally representing the facies of estuarine bars, shoals, and spits. Without providing a detailed description of different facies, we shall emphasize two of the most important structural features of lower Carboniferous cycles.

The first feature is the sharply assymmetric structure of the cycles. In a vast majority of cases, rocks immediately overlying the coal are the most definitely marine for the given cycle; they change upward to rocks which indicate the approach to the shore line. Thus the Çonets cycles include the regressive series of facies, with the transgressive series missing, as a rule. In this they differ sharply -- V.V. Koperina notwithstanding -- from the Pechora Basin Vorkuta cycles. The latter have a symmetric structure with an approximately equal development of the regressive and transgressive components [2, 6].

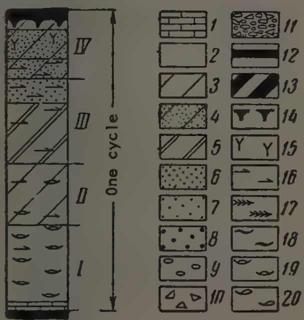

FIGURE 2. Diagram of a sedimentary cycle for lower Carboniferous C_1^3 coal-bearing formation, in the Donbas.

1 -- limestone; 2 -- mudstones, chiefly silty; 3 -- siltstones, fine-to medium-grained; 4 -- siltstones, coarse-grained; 5 -- alternation of siltstones of different grain size; 6 -- sandstone, very fine-grained; 7 -- sandstone fine-grained; 7 -- sandstone fine-grained; 8 -- sandstone, medium-grained; 9 -- pebble beds; 10 - nonrounded mudstone and siltstone inclusions; 11 -- conglomerate; 12 -- coal; 13 -- coal shale; 14 -- fossil Carboniferous soil, "curly"; 15 -- root remains (autochtonous); 16 -- poorly preserved plant remains; 17 -- well-preserved plant remains; 18 -- coarse carbonaceous plant remains (wood fragments); 19 -- fauna; 20 -- indeterminate fauna; I-IV -- segments of the coal accumulation zone (subzones).

In those comparatively rare instances where the coal is separated from the "most marine" rocks above it, by intermediate rocks forming a transgressive sequence, this latter is found to be very much restricted in its thickness, the facies assemblage, and usually in its areal extent. As such, it obviously is determined by local factors. This is illustrated by the data in Table 2.

coast did not remain the same but changed both laterally and with time.

FACIES TYPES OF CYCLES

The nature of the lower, "the most marine," member of a cyclic sequence -- facies

Table 2

Structural asymmetry in the cycles of coal-bearing formation C_1^3

Members	Total Number of Cycles	Average Thickness of a Cycle, in m	Number of Cycles with a Transgressive Series Represented by:		Number of Cycles with Transgressive Series, Relative to the Total No., in %	Thickness of Transgressive Series, Relative to the Total Thickness of Cycles, in %
			Two facies	Three facies		
Upper	324	6	61	8	21	8
Lower	480	5	41	7	10	3

It appears from Table 2 that the assymmetric structure is especially characteristic of the lower, properly coal-bearing, member where the transgressive facies account only for 3% of the total thickness of the cycles. On that basis, we have concluded that if a cycle is begun with the transgression, its facies type can be characterized for the given conditions by the regressive series of facies, as the most complete and the most widely developed.

The second structural feature of the Donets cycles is that hardly any of them contain a gradual transition from open-sea facies to the shallow and then to the embayment, lagoonal, and marsh-lacustrine. Such a transition is effected only over large areas, and only by a gradual change in the facies composition of the cycles, in a definite direction and for a long period of time, through a series of cycles replacing each other in the vertical section. Each individual cycle contains a small number of facies which alternate in a definite sequence.

This sets forth the premises for our classification of the cycles by paragenetic associations of regressive facies. Generally speaking, the alternation of facies in a cycle, regardless of their nature, reflects a transition from segments of subzone I, the farthest away from shore, to the nearest-to-the-shore segments of subzone III and then to the on-shore marsh-lacustrine deposits of subzone IV. It follows that different facies types of cycles characterize their own morphology of the coastal zone of accumulation of coal measures, and that the general aspect of the

of an open sea, shallow sea, embayment, or lagoon -- determines, in most cases, the nature of the remaining components of that cycle. An exception is the wedging-in of fluvial sands which rest with a slight submarine erosional contact on different deposits, from embayment-lagoonal to shallow marine, and which represent a special type of cyclic sequence. Five principal cycle types have been identified, as illustrated in Fig. 3. Each type is characterized by a column showing the percent content of its component facies. Each column is a result of statistical computation, by averaging all of the observed full regressive sequences of the given type. In addition, a typical example of each type section is given (left column - facies; right column -- their representative rocks). Some of the types have several variants of the regressive development, with the approach to the shoreline.

Marine type. This type always contains open-sea facies M, of normal salinity; shallow facies MM (represented mostly by the argillaceous and silty subfacies); and littoral facies MV, affected by the action of waves and of ebb-and-flow currents (Fig. 3, columns 1, 2). These facies correspond to coal-bearing subzones I-III (see Fig. 2). All other facies of the column account for less than 50% of its volume, with the coal beds occurring as a rare exception and never attaining a workable thickness. The average thickness of the entire regressive sequence of this type is 6 m. Fig. 3, column 3 shows a marine type section of exceptional thickness and a comparatively better development of marine facies M. Such a structure is typical

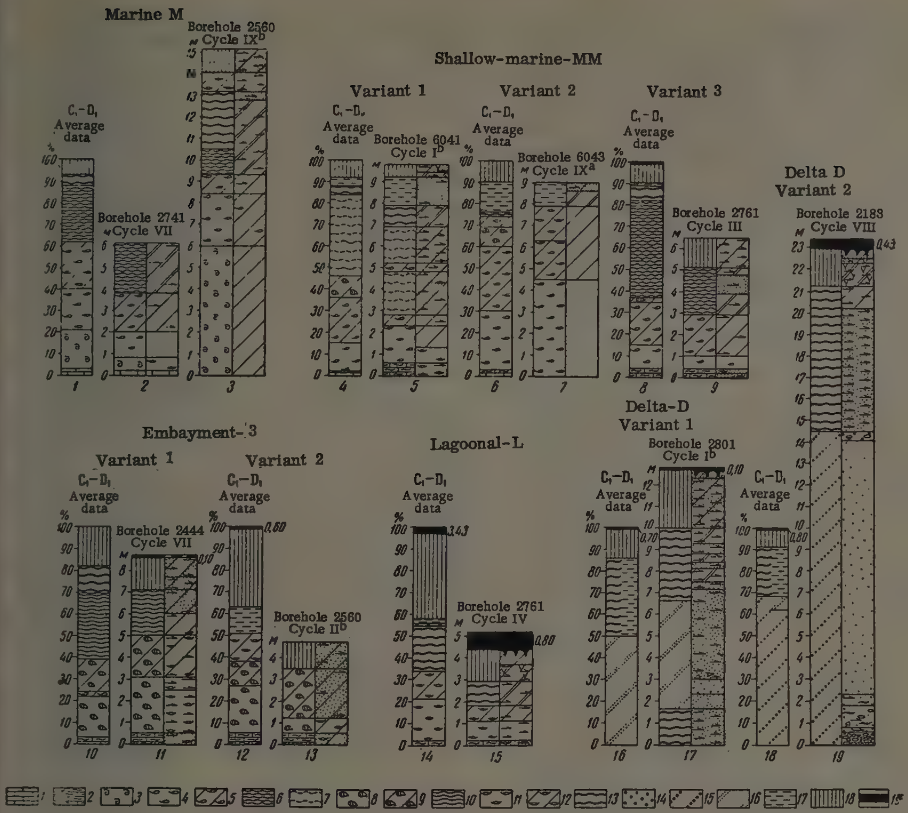

FIGURE 3. Main types of cycles.

1 -- Open-sea carbonate facies MI; 2 -- littoral calcareous siltstone facies (beginning of transgressive stage), MVT; 3 -- open-sea silty-argillaceous facies M; 4 -- shallow sea silty-argillaceous subfacies MMa; 5 -- shallow-sea silty subfacies M^b; 6 -- facies MV of alternating silts and very finely sandy sediments of littoral zone of an open sea; 7 -- facies MVS of homogeneous littoral silts with poorly expressed wave action; 8 -- argillaceous-silty embayment facies Z; 9 -- silty embayment facies ZP; 10 -- silty facies ZV of wave-affected embayment facies ZV; 11 -- silty lagoonal facies ZP; 12 -- silty lagoonal facies LP gravitating toward lacustrine; 13 -- silty to very finely sandy lagoonal facies LV affected by wave action; 14 -- facies LB of sandy bars, spits, shoals; 15 -- submarine delta-type sandy facies D with poorly expressed wave action; 16 -- same with more intensive wave action; 17 -- lacustrine type silty to very finely sandy facies O; 18 -- silty facies BP of marshy coastal lowlands. For explanation of symbols see Figure 2.

of middle Carboniferous coal measures, where similar cycles carry workable coal beds. In the lower Carboniferous, it is an isolated anomalous occurrence.

We associate the wide development of marine cycles with a marine littoral zone which has a leveled shore line and a very gently sloping bottom (near the shore) of the modern shoal type.

Marine cycles are very restricted in the lower Carboniferous where they have a somewhat stunted aspect. Apparently they reflect only isolated stretches of the coast, which in places resembled that type whereas the morphology of the coast as a whole was quite different. These peculiarities in the structure and development of lower Carboniferous marine types appear to have been responsible for their lack of coal beds.

The <u>shallow marine type</u> opens with very fine-grained shallow marine facies MM, with argillaceous subfacies MM^a always at the base, to correspond to subzone I; higher up

in the section, it is replaced by silty facies MM^b, corresponding in this case to subzone II. Littoral subzone III is represented here by coarse-grained siltstones of facies MVS characterized by a very slight wave action. This facies is peculiar only to variant 1 of the shallow marine sequence (Figs. 3-5). The remaining facies of column 4 are rare and are of subordinate character; coal beds have not been observed here. Variant 2 differs by its lack of facies MVS; the lacustrine facies is widely developed, instead (Fig. 3, columns 6, 7). At the same time, the shallow-sea facies is even more widespread, over and above the total of all other facies. Again, no coal beds have been encountered here. Variant 3 is very similar to the marine sequence in its facies and their relationship. At first glance, the only difference between the two appears to be the lack of the open-sea facies. However, a number of additional features such as the fairly common appearance of definitely shallow-water calcareous silts, not present among shallow-marine sediments of the marine-type cycles, as well as the different composition of their concretions, suggest that variant 3 belongs in most instances to specific shallow marine conditions different from those of a littoral marine sequence.

Some light on this problem is shed by the areal distribution of variant 3. In most cases, variant 3 is replaced, over short distances as much as 1 km, by other shallow marine variants. In the transition from the lower, properly coal-bearing member, to the upper and more definitely marine, variant 3 persists over much longer distances, nearly everywhere in the south Donbas district; in the Dobropol'ye district, it is replaced by a marine sequence, thereby displaying a definite kinship with the latter. Characteristically, the shallow marine cycles here are much thicker; they also carry a consistently workable coal bed c_{18}.

The overall thickness of the shallow-water type variants is the same as for the marine type, i.e., 5-7 m.

The broad development of shallow marine cycles can be related to a marine coast, sheltered over long stretches from coastal waves and currents, and apparently with a steeper slope of the bottom, as compared with the conditions of the marine cycles formation. The rare faunal findings, always poorly preserved, give no indication of a change in the salinity of this shallow zone.

The embayment type. Its base is made up of very fine-grained facies Z marked, along with a mixed fauna, by very peculiar, fully pyritized plant remains reminiscent of algae. Higher up in the section, there are first the silts of facies ZP (subzone II), then more silts of facies ZV corresponding to subzone III, with evidence of wave action. The section is culminated in marsh deposits which are much more common here. In the upper member, variant 1 of the embayment sequence carries a workable coal bed of very varied thickness (Fig. 3, columns 10, 11). The average thickness of this cycle type is 7-8 m. Variant 2, which occurs only in the lower member, lacks facies ZV which is replaced by marshy-lacustrine sediments. Coal beds are very rare and noncommercial. Another feature of the cycles of this variant is their small thickness, 4 m on the average.

A broad development of the embayment-type cycle is related either to a transitional type of coast, from the so-called transgression (with a broken coast line) to the leveled-off type; or else to the peripheral (seaward), more open zone of a lagoonal coast.

The lagoonal type is represented by very-fine deposits of lagoonal facies L, with a brackish-water fauna. It changes upward usually to transitional facies LP, and then to the wave-action facies LV and to marsh facies, very commonly with coal beds (Fig. 3, columns 14, 15). Intercalations of calcareous silts are fairly common at the base of this sequence. The lagoonal cycles are, on the average, less than 4 m thick. Their most characteristic features are the exceedingly wide development of marsh formations represented chiefly by typical soils of coal beds, and the high coal content.

The formation of lagoonal cycles is naturally related to a closed type transgressive coast, with a broken coast line and a wide development of lagoons.

Cycles marked by the presence of river loads (submarine deltas). Sandy deposits of the river-load facies D are represented in the lower member almost exclusively by very fine to fine varieties of subfacies D^a, with evidence of a slight fluvial action accompanied by a considerable reworking of the sediments in the wave-action zone. They rest, with a slight and shallow erosional contact on lagoonal-lacustrine-marsh deposits. Their thickness is varied, about 3 m on the average. They are overlain by about the same thickness of a lagoonal-lacustrine-marsh sequence, with the marsh facies of a comparatively subordinate nature (Fig. 3, columns 16, 17). There are some noncommercial coal beds. The coarser sandstones with conglomerate beds at their base -- subfacies D -- are associated chiefly with the upper member. They rest chiefly on shallow marine deposits, with a sharp erosional contact of a greater magnitude. Their average thickness is as much as 9 m. They, too,

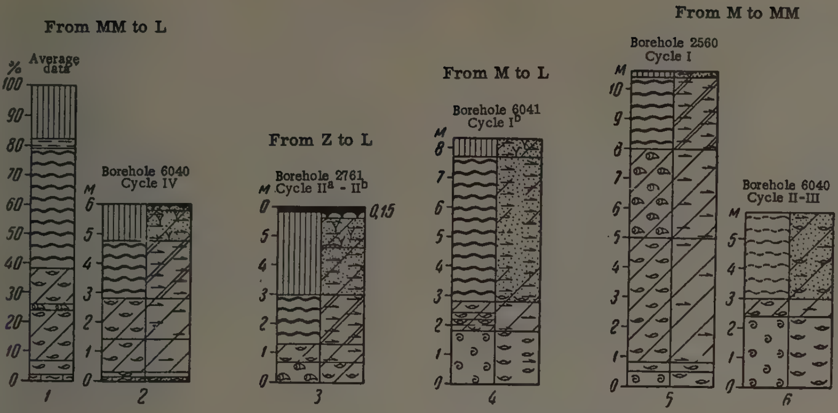

FIGURE 4. Transitional types of cycles.

For symbols to lithologic columns see Figure 2; to facies columns, Figure 3.

change upward to lagoonal-lacustrine and then to marsh facies (Fig. 3, columns 18, 19). In isolated instances, they carry coal beds of a very varied thickness.

The appearance of river loads, especially of the coarser subfacies D^b, is constantly accompanied by an increase in the thickness of the regressive sequence. The latter is about 10 m thick, for subfacies D^a, and 15-18 m for subfacies D^b, on the average.

The wide development of cycles marked by the participation of river loads is related to the approach of an estuary, and possibly to a rejuvenation of the relief at the source.

In two instances the upper member was found to contain, besides the above-named types, alluvial sandstones resting with an erosional contact on lacustrine-marsh or deltaic deposits, and changing upward back to the lacustrine-marsh.

Transitional types. Besides the main facies types -- each characterized by a definite predominance of marine, shallow marine, embayment, lagoonal, or deltaic group of facies -- there are types with a transition from one facies group to another, less definitely marine. Instances of such cycles are given in Fig. 4, with the following transitions: a) from shallow to lagoonal facies (Fig. 4, columns 1, 2); b) from embayment to lagoonal (column 3); c) from marine to lagoonal (column 4); and d) from marine to shallow marine (column 5). In addition, there are some similarly organized cycles showing a transition from marine and shallow marine facies to the embayment. Cycles transitional from embayment and shallow marine facies to the lagoonal carry mostly noncommercial coal beds, with very rare inconsistent commercial beds. The remaining types are practically barren.

The broad development of transitional cycles suggests a remodeling of the coast, accompanied by a restriction of the area formerly occupied by a facies group, such as the lagoonal, and a landward migration of more definitely marine facies, such as the shallow marine. As a result, two facies groups instead of just one are found within the coal-accumulation zone of a single cycle such as the lagoonal in the immediate vicinity of the shore and the shallow marine some distance away from it.

DISTRIBUTION OF DIFFERENT TYPES OF CYCLES THROUGHOUT THE COAL-BEARING FORMATION, AND FACIES CONDITIONS OF ITS ACCUMULATION

In a most general way, the distributions of the several cycles, both vertically and laterally, can be characterized by the relative content of each type in the upper and lower members, by individual districts (Table 3), as has previously been done for the facies (in Table 1). It appears that transitional types are everywhere poorly developed. The only exception is the type transitional from a shallow sea to a lagoon, which is as common as the main types. A comparison of the lower and upper members shows a growing importance of marine and

shallow marine cycles in the latter, with a sharp curtailment of lagoonal cycles. Presented this way, the differences between the two members show better than in the correlation of their facies composition.

A correlation of individual districts reveal that, for the lower member, districts with a high content of marsh facies -- including the coal beds -- such as the Mezhavaya-Volch'ya prospect and south Donbas, are also marked by a much higher content of the lagoonal-type cycles. Furthermore, there are substantial differences between the Dobropol'ye and Karavannaya-prospect districts. With a similar content for both of the lagoonal and shallow marine facies -- which corresponds to their approximately same coal content coefficient -- the Dobropol'ye district has better developed marine and embayment types as well as a type, transitional from an open to a shallow sea. The higher content of marine cycles in the Dobropol'ye district is even better expressed in the upper member.

Our profiles of the cycles' distribution within the members demonstrate that some of the types persist for a long time and spread all over the area, whereas some others are more restricted in time, and still others appear only sporadically, are laterally inconsistent, and are replaced by other types.

The general development of cycles belonging to one of the main types, and their subsequent persistance over a long period of time, suggest a stable profile and a relative permanence of the morphologic aspect of the coast, also represented by a corresponding type. We believe, therefore, that stable and generally developed cycle types determine the facies environment of a sedimentary process and the corresponding type of the coast with its landscape.

Table 3

Relationship of different cycles within coal-bearing formations C_1^3 (in % of total thickness of members)

Districts	Lower member				Upper member			
	Dobropol'skiy	Mezhevaya-Volch'ya prospect	South Donbas	Karavannaya prospect	Dobropol'skiy	Mezhevaya-Volch'ya prospect	South Donbas	Karavannaya prospect
Types of cycles	Borehole 2560	Borehole 2183	Borehole 2801	Borehole 6040 + 6041	Borehole 2560	Borehole 2183	Borehole 2824	Borehole 6043 + 6044 + 3376
Marine	5.4	—	—	0.9	16.5	6.7	4.6	3.1
Shallow marine	20.0	19.4	9.2	20.7	32.7	21.5	45.8	46.6
Embayment	9.0	2.8	0.6	5.0	8.0	13.0	10.1	5.9
Lagoonal	18.5	46.7	45.4	21.1	3.0	6.1	15.6	10.0
River loads	13.5	4.9	26.4	17.5	12.7	13.8	1.8	7.3
Fluvial	1.5	1.5	—	—	3.9	—	—	4.4
Transition from open to shallow sea	9.5	6.0	6.4	2.4	4.4	3.5	4.7	8.8
from sea to embayment	—	—	—	2.9	2.6	4.2	1.8	—
from sea to lagoon	3.5	3.4	3.7	1.9	—	10.0	—	—
from shallow sea to embayment	1.7	0.9	—	—	8.3	4.2	5.8	6.5
from shallow sea to lagoon	15.6	6.3	4.7	26.5	6.1	10.7	8.7	7.4
from embayment to lagoon	1.7	8.2	3.7	1.0	1.9	6.2	1.1	—

The lower member was subject to a general and repeated development of lagoonal cycles; in central districts (Mezhevaya-Volch'ya prospect; south Donbas), they persisted during the entire period of its formation. Accordingly, that was the period of a lagoonal environment. It was featured by a frequent, albeit local, appearance of deltaic cycles (with river loads) along with the lagoonal. Most commonly, the rivers carried their load to the south Donbas district.

The second feature of a lagoonal environment was the instability of submarine cumu-

lative formations which separated the lagoons. The bars and spits apparently were not large, and subject to frequent breaking-up and rebuilding, seldom passing into fossil state. This is suggested by the poor development of sandy facies of shoals and bars, their areal inconsistency, and a frequent presence at the base of lagoonal cycles, of thin intercalations of coarse-grained calcareous siltstones with an allochtonous littoral marine fauna.

As a result, along with the main lagoonal type of cycles, a subordinate transitional type participated in the build-up of the lagoonal environment -- that of a shallow sea -- lagoon change, with the shallow marine cycle type locally present. The exceedingly high content of this shallow sea -- lagoon transition type in the Karavannaya-prospect district, where it continually alternates with the lagoonal type, suggests a more general effect of regional factors. The most probable is a slight steepening of the littoral shelf, which brought about a narrowing of the lagoonal-facies zone and a corresponding approach to the shore of shallow marine facies.

During the comparatively brief periods of "major transgressions," which separated the large cycles (of third order), shallow-marine cycles were generally developed, and the lagoonal environment changed to the shallow-marine, definitely subordinate, on the whole.

These features of a lagoonal environment prevailing in the lower Carboniferous period of coal accumulation in the Donets Basin are in sharp contrast with a lagoonal environment of the Permian coal accumulation in the Pechora Basin. According to the work of G. A. Ivanov [2], A. V. Makedonov [5, 6], and others, the lagoonal conditions determining the commercial coal-bearing capacity of the Vorkuta deposit are expressed exclusively by marsh and lagoonal facies with a wide development of bar facies. The remains of bars are present in every cycle. The so-called "beyond-the-bar" facies take no part in the productive interval of the section.

Shallow-marine conditions prevailed while the upper member was being formed. The corresponding cycles here were interspersed with deltaic cycles consisting mostly of coarse-grained sediments of subfacies D^b.

Marine conditions were established during the brief periods of major transgressions; they were most stable in the west in the Dobropol'ye district. The lagoonal conditions persisted only at the very base of the upper member. Unlike the marine conditions, they were most stable in the east in the Karavannaya-prospect district, where they witnessed the appearance of fluvial facies. Both the marine and lagoonal conditions are strictly subordinate in the upper member.

The embayment type of cycles does not form an environment of its own. Most commonly it participates as an element of the shallow marine and marine environment, in the upper member; and as a component of the lagoonal types, in the lower. In the upper member, the embayment and lagoonal types reflect the relict forms of the coast, persisting from the predominant lagoonal environment of the lower member. On the contrary, the much wider development of marine cycles in the upper member is related to substantial changes in the coast type and reflects a new trend in its development as expressed in the recurrent leveling-off and in the establishment of a short-lived marine environment. This trend is most definite in the Dobropol'ye district.

FACIES ENVIRONMENTS AND THE COAL-BEARING CAPACITY

The test holes which we studied and which illustrate the entire coal-bearing formation C_1^3 in the four districts, have penetrated in 125 instances, noncommercial beds (less than 0.45 m thick), and in 34 instances, commercial beds. Their distribution by the cycle types is given in Table 4.

In a vast majority of cases the coal beds -- including the commercial -- in the lower member correspond to the lagoonal cycles, the principal type of the prevailing lagoonal conditions. Isolated commercial beds related to other cycles are of the inconsistent type, workable only locally. An exception is the shallow-marine cycle with a consistently commercial coal bed c_{18}. However, as has been stated above, we have here an instance of a shallow-marine type developing according to variant 3, which gravitates toward the marine type.

In the upper member, isolated commercial coal beds occur in cycles marked by the participation of deltas, and in fluvial cycles. The local development of these cycles predetermines the inconsistency of the involved coal beds which rapidly lose their commercial nature. Not a single commercial coal bed is involved in the shallow-marine cycles, despite their local prevalence.

Thus, although the lack of commercial beds among the marine and embayment types may be possibly explained by an inadequate development of the latter (which is undoubtedly true as far as the marine type is concerned), the shallow-marine type -- even where it reflects a predominance of shallow-marine

Table 4

Distribution of coal beds by cycles of different types

Types of cycles	Lower member		Upper member	
	commercial	noncommercial	commercial	noncommercial
Marine, M	—	—	—	1
Shallow marine, MM	3	4	—	3
Embayment, Z	—	3	1	—
Lagoonal, L	21	67	—	5
River load, D (deltaic)	1	5	2	2
River, R	—	—	2	—
Transitional:				
M — MM	—	2	—	—
M — L	—	1	—	1
MM — Z	—	—	—	1
MM — L	3	16	—	6
Z — L	1	8	—	—
Total	29	106	5	19

conditions -- is especially unfavorable for coal accumulation.

On the other hand, the association of coal and the lagoonal conditions, in the lower member, is so definite as to stand out even in the correlation of individual intervals of the coal-bearing formation. Thus a correlation of intervals between limestones $C_1 - C_2^1$; $C_2^1 - C_4^3$; $C_4^3 - C_5$ (within the lower member) and interval $C_5 - D_1$ corresponding to the upper member, as revealed by the test holes, has demonstrated a good correspondence between the contents of lagoonal cycles and marsh facies in the coal-bearing formation (Fig. 5). A similarly good correspondence (Fig. 6) is obtained from the correlation for the same intervals, of their marsh-facies content and their coefficient of the coal-bearing capacity within a lagoonal environment. It appears that the marsh-facies content is directly proportionate to the lagoonal-cycles content, whereas the coal-bearing capacity of the lagoonal cycles is, in turn, directly proportionate to the marsh-facies content.

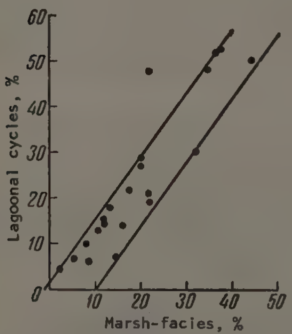

FIGURE 5. Relationship between the lagoonal cycles content and that of marsh facies.

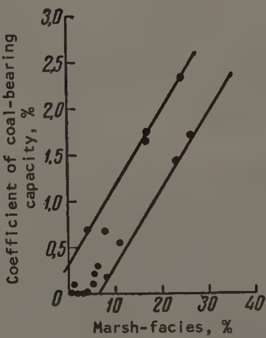

FIGURE 6. Relationship between the coal-bearing capacity coefficient and the content of marsh facies in a lagoonal environment.

CONCLUSIONS

Our data show that a more concrete concept of the productivity or the lack of it for different types of coal measures, and of the causes of the more or less narrow range of changes in the coal-bearing capacity, is impossible without an analysis of the facies environments and of their changes, both laterally and with time. Facies environments characterize the prevailing coast type in the confines of the coal-bearing zone. The changes in the coast type, both lateral and vertical, bring about substantial and uniform changes in the coal-bearing capacity, even under the more or less stable geotectonic conditions of the basin, its general paleogeography, and its climate. For this reason, all those features characteristic of the productive section type will persist, as a rule, only within a definite environment. For instance, G. A. Ivanov's thesis [3] to the effect that the thickest cycles appears to be valid for a uniform lagoonal environment of the Vorkuta deposits. It is invalid in the more facies-diversified environment of the Donets Basin lower Carboniferous, where the coal-bearing capacity is associated with lagoonal cycles of the maximum thickness (as compared with other types).

Thus, a differentiation and classification of facies environments must precede all attempts at the establishment of any pattern in the occurrence of coal in coal measures of any basin.

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LITHOLOGIC- FACIES DESCRIPTION OF THE LOWER CARBONIFEROUS CARBONATE SERIES ON THE NORTHERN SLOPE OF THE UKRAINIAN CRYSTALLINE MASSIF[1]

by

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Tournaisian and lower Visean deposits in the northern slope of the Ukrainian crystalline massif are described. It is established that the carbonate sequence wedges out gradually, going westward, instead of being fully replaced by terrigenous sediments. For this reason, terrigenous and carbonate deposits of the western Donbas districts, formerly correlated with the Donbas C_1^Vf zone, are really correlative with the C_1^Vg zone.

* * * * *

In the recent prospecting by the Artemuglegeologiya and Ukrugleogeologiya Trusts, and the Ukrainian Geological Administration, Tournaisian and lower Visean deposits were penetrated by a number of boreholes and traced along the northern slope of the Ukrainian crystalline massif as far west as the Novo-Moskovsk district. The correlation and stratigraphic differentiation was made by D. Ye. Ayzenverg, N. Ye. Brazhnikova, M. V. Yartseva, F.M. Dyssa, P.G. Nesterenko, V.I. Pogodina, and A.Z. Shirokov [1-6]. They came to the conclusion that Tournaisian and most lower Visean carbonate deposits, become thin sharply, going westward; are replaced by terrigenous deposits, and wedge out completely in the Novo-Moskovsk district. The carbonates of zone C_1^Vf begin to get replaced by terrestrial deposits as distant as in the Mezhevaya area. This opinion is shared by S.V. Trofimov [16], A.P. Stukalo, and V.N. Stovpovoy [14, 15].

V.Z. Yershov holds a different view. At the Second All-Union Coal Conference, he denied the possibility of a change of lower Carboniferous limestones to terrigenous deposits, in the west of the greater Donbas. P.L. Shul'ga recently arrived at the same conclusion [18].

Such contradictory views are explained, in our opinion, by the lack of detailed lithologic descriptions of these deposits, without which one cannot obtain a comprehensive idea of their origin and of the geologic history of the Tournaisian and lower Visean, of this region.

LITHOLOGY OF THE CARBONATE SEQUENCE

In Donbas outcrops, the Carboniferous begins with thick Tournaisian and lower Visean carbonates. They were studied by many geologists [7-12] and they have been divided into a number of faunal and lithologic zones. The most detailed differentiation of these rocks was made by A.P. Rotay [9].

Our own studies have shown that A.P. Rotay's stratigraphic zones are traceable far to the west where they maintain their general structural features, although decreasing in thickness, until they gradually wedge out.

In south Donbas, lower Carboniferous carbonate rocks are underlain by Devonian sedimentary and extrusive-sedimentary formations. To the west, they rest upon Precambrian extrusives and metamorphics, represented mostly by granites, their migmatites, granite-gneisses, and gneisses. In the area of Bogatyr' Village, they rest upon amphibole schists.

A weathering crust is developed on Precambrian rocks. The weathering product is an argillaceous hydromicaceous-kaolinite substance with a large number of angular quartz grains, fully or nearly fully weathered orthoclase and plagioclase grains, and grains of microcline and of weathered or chloritized biotite and muscovite.

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[1]Litologo-fatsial'naya kharakteristika karbonatnoy tolshchi nizhnego karbona severnogo sklona ukrainskogo kristallicheskogo massiva.

The base of the carbonate sequence almost everywhere carries a bed of poorly sorted sandstone, 7 to 10 m thick. Its lower part contains numerous quartz and siliceous fragments, more or less rounded, as much as several centimeters in size. In the upper part, the sandstone is finer grained, consisting of ill-sorted and differentially rounded grains of quartz, orthoclase, plagioclase, microcline, and quartzite fragments, with a siliceous to carbonate cement. The sandstone is commonly overlain by a shale bed, as much as 3 m thick, carrying as much as 30 percent angular, ill-sorted quartz grains in its lower part, whereas its upper part is made up almost fully of beidellite and hydromica scales. Locally, there is a gradual transition from shale to the overlying carbonate rocks.

The above-named features of arenaceous and argillaceous deposits underlying lower Carboniferous carbonate rocks suggest that they were littoral marine sediments deposited under the conditions of a progressive Tournaisian transgression. The overlying carbonates are a continuation of the same transgressive sequence. On the basis of these facts, we concur with D. Ye. Ayzenverg's opinion that these rocks are Carboniferous rather than Devonian.

The lower member of the Donbas sequence is represented by micrograined, locally dolomitic limestones of zone $C_1^t a$, interbedded with shales. Their maximum thickness of 220 m has been observed in the Novo-Troitsk area. In the area under study, zone $C_1^t a$ is correlated with massive dark-gray, chiefly microgranular to cryptogranular, slightly dolomitic limestones in the lower portion of the carbonate sequence, in the area of the Bogatyr' and Alefirovka Villages. These limestones are locally interbedded with dark-gray to greenish-gray shales and are made up of a pelitomorphous and micrograined calcite; in other places, the recrystallized segments are associated with aggregates of rhombohedrons and isometric grains of dolomite. Organic remains are extremely rare, being represented by valves and whole ostracod shells. Much less common are recrystallized fragments of brachiopod shells. Pyrite is present everywhere in large spots, and disseminated in fine grains throughout the rock. There are isolated quartz grains, angular to rounded, 0.25 to 1.0 m. The micrograined and pelitomorphous limestones are 8.7 m thick in the Bogatyr' area, and 7.0 m in the Alefirovka area. Their correlatives have not been observed west of there.

Carbonate rocks of zone $C_1^t a$ appear to have been deposited, in the area under study, in a shallow marine embayment, with the main marine basin to the east.

Deposits of zone $C_1 b$ are developed along the northern slope of the Ukrainian crystalline massif. In the Novo-Troitsk area, they are represented by dolomites, dolomitic and microgranular limestones, with montmorillonite intercalations in the upper part of the sequence [8]. The $C_1^t b$ zone is 80 m thick, decreasing to the northwest, where it is 32 m thick in the Bogatyr' area, 60 m in the Novo-Pavlovka, 36 m in the Alefirovka, 12 to 18 m west of Pavlograd, and 15.8 m at Znamenka. A little west of Novo-Moskovsk this zone completely wedges out.

As in the Novo-Troitsk area, it is represented by dolomites and dolomitic micrograined limestones, massive, gray to light-gray and yellow gray, less commonly light yellow, brownish to brown-gray. Under the microscope, the finely granular limestones are seen to consist of calcite and dolomite grains. The dolomite is usually represented by isometric grains and rhombohedrons, 0.01 to 0.05 mm, less commonly to as much as 0.1 mm, with refraction indices $\omega = 1.680$, $\epsilon = 1.535$. Some of the dolomite rhombohedrons exhibit a zonal structure. The calcite grains are somewhat larger; their maximum size, however, is never more than 2.0 mm. Refraction indices for calcite are $\omega = 1.66$, $\epsilon = 1.48$.

Both the calcite and dolomite are unevenly distributed. In some areas the rock is all dolomite, with isolated crystals and veinlets of calcite; in other places calcite and dolomite grains are present in the same amount. Less common are beds of pure dolomite and limestone. Pyrite is present nearly everywhere in dispersed rounded grains. Veins and crystals of barite, as much as 2 mm, are present locally, in places, containing inclusions of carbonate crystals.

Refraction indices for barite are: $\gamma = 1.648$, $\beta = 1.637$, $\alpha = 1.636$. There are a few isolated quartz grains, as much as 0.25 mm, and small plant shreds. No faunal remains have been found.

The lithology of zone $C_1^t b$ is somewhat different in the extreme west, in the Znamenka area, where chiefly limestones, more or less crystallized and slightly dolomitic, are present, locally interbedded with shales. Macroscopically, the limestones are gray to light gray and yellow-gray, usually with a distinct horizontal lamination of sandy-silty to argillaceous material.

These limestones are also marked by a higher pyrite content, in spotty aggregates of various sizes and in individual crystals evenly dispersed throughout the rock. The pyrite inclusions are commonly associated with the argillaceous laminae.

The C_1^tb section is terminated by microgranular, slightly dolomitic limestones with a small amount of organic remains, mostly of ostracods and recrystallized brachiopod fragments.

As revealed by the chemical analyses, the amount of insoluble residue in zone C_1^tb increases to the west, to reach its maximum (11.6-14.42%) in the area of its wedging out (Znamenka). The dolomite content decreases in the same direction, and upward through the section. For instance, in borehole No. 169, the $MgCO_3$ content in interval 223-230 m changes from 22.3 to 44.8%, and the $CaCO_3$ content changes from 76.9 to 50.3%.

Carbonate rocks of zone C b represent deposits of a rather shallow-marine basin with a higher carbonate salt concentration which hampered the development of organisms and determined the exclusively chemical precipitation of calcite and dolomite.

The higher salinity of the basin apparently was a result of its shallow depth and of an arid climate of that time. The difference in dolomitization throughout the basin is explained by local conditions. Most probably, it was a function of the depth, differential heating, and uneven influx of continental waters. The extreme western areas presented at that time a small sheltered embayment. Because of the influx of continental fresh waters carrying terrigenous material, calcium carbonate was the main precipitate, with an addition of silty clay particles. The upward decrease in the dolomite content in zone C_1^tb suggests that salinity of the basin decreased gradually until it became normal everywhere during the deposition of zone C_1^tc. The reason for the frequent occurrence of dolomite in spots, intercalations, and lenses is a diagenetic redistribution of it in the rock. N.M. Strakhov [13] assigns such dolomitic rocks to a sedimentary-diagenetic type.

A.P. Sklyar voiced an opinion [11, 12] that the dolomitization of lower Carboniferous carbonate deposits was caused exclusively by the intruding pink-gray aplitic granites. This view is contradicted by the fact that dolomitic rocks nearly everywhere rest upon almost nondolomitic limestones or else directly upon weathered Precambrian rocks.

According to N.D. Reshetnyak [8], zone C_1^tc is represented in its Yelenovka, Styla and Kal'mius sections by porphyrylike limestones with a rich fauna of brachiopods, corals, and foraminifera, and with chert lumps in its middle part. This zone is the thickest in the Novo-Troitsk area, where it is 38 m. It thins down, going northwest, until it wedges out completely in the Kulebovka area.

Throughout the area under study, the C_1^tc limestones are light gray to yellow-gray, locally almost white to brownish gray, finely crystalline, carrying phenocrystlike inclusions of coarse-crystalline calcite with inclusions of brachiopod shells, crinoid stems, and corals. In the Bogatyr', Novo-Pavlovka, and Alefirovka areas, the lower part of the section carries argillaceous intercalations and chunks made up of beidellite scales. The upper part commonly contains rounded chunks of black to gray chert, as much as 3 cm in diameter.

The C_1^tc limestones are 55-99% organic remains in an organogenous matrix, and are represented by clots and tubular inclusions of algae. Present to a smaller extent are thick-walled foraminifera, large ostracods, crinoid segments, and brachiopod-shell fragments. The organogenous material is cemented by a differentially crystallized calcite substance. The calcite crystals vary in size from 1.0 to 3.0 mm.

The C_1^tc limestones commonly carry quartz grains of irregular outlines, and locally barite and pyrite.

The limestones are usually pure. Only in the extreme west, in the Kulebovka area, are they enriched by terrigenous material and interbedded with shales. The amount of brachiopod-shell fragments and crinoid segments in rocks increases in the same direction, from 5 to 7% in the east to 15 to 20% in the west (Znamenka). This suggests a more shallow sedimentation in the west. In their chemical composition the C_1^tc limestones are fairly homogeneous.

In most samples the insoluble residue does not exceed 1%. The MgO content is constantly less than 1% which points to a low degree of dolomitization of these rocks. The thermal curves for the limestones also demonstrate a chiefly calcite content.

The lithology and faunal content of zone C_1^tc confirm that these limestones were deposited in a sea with a salinity normal for that time, which promoted an intensive development of marine organisms. The general presence of a considerable amount of algal remains suggests a depth of not more than 50 m.

Zone C_1^td, in southern Donbas districts is made up chiefly of finely granular limestones, with remains of ostracods, algae, and other organisms in its upper part, and calcareous sandstone lentils and stigmaria molds [8]. Its thickness in the Novo-Troitsk area is 25 to 25 m. The presence of C_1^td limestones has not been established farther west along the crystalline massif slope. Furthermore, evidence of weathering was

found in this zone in the Pavlograd area and west of there, as it had been found earlier by D. Ye. Ayzenverg [1-4], for the Novo-Moskovsk area.

The above-described Tournaisian deposits on the northern slope of the Ukrainian crystalline massif are everywhere overlain by lower Visean deposits. In the southern Donbas districts, (Kal'mius River, Styla and Novo-Troitsk Villages), the lower Visean sequence begins with zone $C_1^V a$, as much as 5 m thick. According to N. D. Reshetnyak [8], the latter is everywhere represented by algal and foraminiferal-algal, commonly argillaceous, limestones. Limestones of zones $C_1^V b$ and $C_1^V c$ are similar in their lithology.

In more western areas of the northern slope of the Ukrainian massif, the deposits of these first three lower Visean zones cannot be distinguished, lithologically; accordingly, we regard them as a single bed $C_1^V a$-c. Its rocks rest in the east directly on unaltered deposits of the $C_1^t c$ zone; west of Pavlograd they rest on weathered limestones of that zone; west of Kulebovka, directly on Precambrian crystallines. Its maximum thickness, about 45 m, has been observed at Novo-Troitsk; it gradually decreases going northwest, until it wedges out completely somewhat west of the Kil'chen' River valley.

East of Alefirovka, there are chiefly organogenous dark-gray, cryptogranular to finely granular massive limestones, interbedded in their lower part with dark-gray calcareous shales. Organic remains predominate in the limestones, chiefly of algae, thick-walled foraminifera, and an undeterminable organic detritus. Present to a smaller extent are ostracods, brachiopod- and pelecypod-shell fragments, shreds of bryozoa, and crinoid segments. This organogenous material is cemented with a cryptogranular to finely granular calcite matrix everywhere permeated with an argillaceous substance. Locally the limestones have a knotty structure, because of rounded nodules, as much as 5 mm, of pure microgranular foraminiferal limestone. In the Bogatyr' area, the upper part of bed $C_1^V a$-c contains, in places, nodules of dark-gray chert with cavities filled with pyrite, calcite, and galena.

Considerable changes take place in the structure of this bed, west of Alefirovka: terrigenous sediments, 0.2 to 1.5 m thick occur here along with the carbonate.

The carbonate rocks are made up of gray, massive, arenaceous, commonly horizontally-stratified limestones with 3-70% organic remains, mostly algae, foraminifera, and indeterminate detritus. Less numerous are fragments of brachiopods, pelecypods, crinoid segments, ostracods, and echinoid spicules. Many samples contain organic remains. The bulk of the rock is represented by a cryptogranular to finely granular ground mass of calcite grains, 0.05 to 0.08 mm, less commonly 0.1 to 0.25 mm, isometric to serrate in form. Pyrite is present everywhere in large spots and disseminated in fine crystals. Barite occurs here and there. Traces of silification have been observed in some of the limestone beds in the middle part of the sequence, in the Pavlograd area. These silicified segments usually carry much secondary quartz, which fills the hollows of shells and corals.

Carbonate rocks of bed $C_1^V a$-c are marked by their higher insoluble residue (commonly more than 20-25%) and by their predominately calcitic composition. Thermal study confirms the results of the chemical analyses.

The principal terrigenous rocks are dark gray, less commonly greenish-gray, commonly calcareous shales with broken and whole shells of brachiopods and pelecypods, and with plant detritus. The rocks are usually massive, although nearly always marked by thin slaty cleavage. The argillaceous rocks consist of scales of beidellite and hydromica, schistose to matted fibrous in structure. Locally, the shales carry small plant shreds and a sizable amount of fine iron-sulfide grains. The results of chemical analyses on individual samples show about 60% SiO_2, 10-20% Al_2O_3, 5-6% Fe_2O_3, 1-2% CaO, and 1-3% MgO.

In the extreme west, fine-grained, more or less calcareous siltstones with fine horizontal partings and with remains of a marine fauna and plant shreds, occur besides the shales.

Most probably, rocks of bed $C_1^V a$-c were deposited under the conditions of a shallow sea, with adequate aeration and normal salinity. Deeper stretches were located in the south Donbas, to provide more favorable conditions for the precipitation of chiefly organogenous calcium carbonate. West of Alefirovka, the large influx of terrigenous material hampered an abundant development of organisms; accordingly, chiefly chemical precipitation of calcium carbonate took place there. The presence of thin horizontal partings and of a considerable amount of pyrite in rocks of the extreme west suggests a comparative quiescence and a poor aeration of the water body, which, in turn, did not encourage the growth of marine organisms. A small marine embayment appears to have existed at that time, in the Novo-Moskovsk area.

In the Novo-Troitsk area, zone $C_1^V d$ is

represented chiefly by foraminiferal limestones and is marked by a considerable amount of nodules, lenses, and intercalations of chert. The overlying zone C_1^Ve, according to N. D. Reshetnyak [8], consists exclusively of siliceous rocks with sponge spicules and radiolaria. Siliceous limestones with siliceous intercalations and nodules are traceable farther west. Their thickness, however, decreases sharply, being only 4.5 m west of Pavlograd. Corresponding beds have not been found in the Novo-Moskovsk area.

Limestones of this bed are dark gray, locally with a distinct horizontal or gently wavy stratification which is formed by an alternation of 5-10 mm thick siliceous lentils and dark argillaceous varieties.

The limestones consist chiefly of thick-walled foraminifera tests, tubular and nodular algae, coarse fragments of brachiopods and pelecypod shells, crinoid segments, ostracod valves, and indeterminate organic detritus. Locally there are sponge spicules filled with calcite and chert.

There are some lentils of clastic limestones consisting of rounded light-colored fragments, 0.3 to 1.0 cm, cemented with a dark-gray argillaceous limestone. These beds appear to have been formed under the conditions of intensive wave action, with the erosion of earlier and hardened deposits.

Chemical analyses and thermal study show that limestones of bed C_1^Vd-e are essentially calcitic, locally with an addition of manganese carbonate.

The silicification of the section is uneven and grows more intensive upward. The siliceous substance occurs in dispersed microscopic inclusions, large nodules, lenses, and intercalations. The microscopic siliceous inclusions are associated, in most places, with the crystallized segments and are usually represented by chalcedony. The latter locally is precipitated in the organic remains. Siliceous nodules are fairly common; they are usually isometric in form, dark to dark gray, almost black, as much as 5 cm in diameter. Some of the larger nodules consist of smaller siliceous bodies cemented with a siliceous substance or else by carbonate. At times, they have a banded structure formed by an alternation of light and dark bands, the latter enriched by pyrite. Such nodules are commonly observed to have cavities filled with beidellite, pyrite, and sphalerite, with calcite veins.

Siliceous layers occur mostly in the upper part of siliceous limestones, 0.05 to 1.40 m thick. Macroscopically, it is a gray to light gray and almost white massive rock, cryptocrystalline. Locally it carries black siliceous nodules with a fringe of fine pyrite grains. Some of the beds display thin horizontal stratification caused by partings, 1 to 2 mm thick, enriched by argillaceous material, fine plant detritus, and pyrite. The siliceous layers are made up of chalcedony grains, as much as 0.01 mm, cemented by amorphous silica.

Their composition, as determined by chemical analysis, is as follows: SiO_2, 96%; Fe_2O_3, 1.27%; CaO, 0.45%; MgO, 0.66%.

The thermal curve of this siliceous rock shows a small endothermic effect with a maximum at 565°C, and a slight exothermal effect at 1065°C. Both the chemical and thermal analyses confirm the siliceous nature of these rocks.

Here and there, carbonate rocks of bed C_1^Vd contain brown concretions of pelitomorphous brownish-yellow oligonite, broken by fractures which are filled with fine-grained calcite; also limestone layers somewhat enriched by manganese carbonate. In the Alefirovka area, the upper part of this bed contains layers of fine-grained calcareous sandstone and sandy limestone.

Some students believe that limestones enriched by the siliceous substances, as well as siliceous rocks of bed C_1^Vd-e, are deep-water formations [8]. According to our own observations, the rocks of this bed were deposited in an environment more shallow than of the underlying C_1^Va-c bed. This is suggested by the large amount of coarse clastic organogenous material in the limestones, by the evidence of erosion of the underlying beds, and by the presence of plant detritus and sandstone and sandy limestones.

We regard these siliceous layers as sedimentary formations deposited in a shallow marine basin, under normal hydrodynamic conditions. The problem of the deposition of silica is much more difficult. A detailed microscopic study of limestones and siliceous rocks of bed C_1^Vd-e has shown that organic remains with a siliceous skeleton are exceedingly rare. It is therefore hardly to be supposed that all of the silica was of organic origin. Most probably, much of it was deposited in colloid flakes. As to the individual nodules and segregates of silica scattered throughout the rock, they appear to be typical diagenetic formations.

Zone C_1^Vf terminates the Tournaisian - lower Visean carbonate sequence. In eastern districts (Yelenovka, Styla, and Kal'mius sections), these deposits are represented by micrograined detrital limestones with a mixed fauna. According to N. D. Reshetnyak [8],

limestones in the lower part of this zone commonly carry chert inclusions. Its middle part is marked by mudeaters' tracks and by sutural-stylolite surfaces. In the Yelenovka section, this interval contains calcareous sandstones and breccialike limestones.

The maximum thickness of this zone (85 to 90 m) was observed in the east, in the Novo-Troitsk area. In the west, it is sharply curtailed, to 11.0 m at Bogatyr', and 2.25 m at Novo-Pavlovka. Farther on, it wedges out completely, with the siliceous bed C_1^Vd-e overlain directly by arenaceous and argillaceous deposits of zone C_1^Vg.

In the area under study, rocks of zone C_1^Vf are represented by gray to light gray limestones, cryptocrystalline to granular, with fragments of brachiopod shells, 1.5 x 0.5 mm, crinoid segments, and isolated corals. The faunal remains are commonly recrystallized and form inclusions of coarsely crystalline calcite.

Limestones of bed C_1^Vf are chiefly organogenous-clastic; they contain 45 to 60% remains of organisms, cemented with an unevenly crystallized carbonate matrix.

Its fauna is represented chiefly by brachiopod and crinoid remains, with foraminifera and algae accounting for 2 to 5% of rock volume; in addition, there is a large amount of small ostracod valves and other indeterminate detritus.

The C_1^Vf limestones are characterized by their low insoluble residue (as much as 2.12%), their high Cao content (51.96%), and the insignificant MgO (0.56%), MnO (0.34%), and FeO (0.28%) content.

The predominantly calcitic content of these limestones is confirmed by thermal curves.

Judging from their lithology and distribution, deposits of zone C_1^Vf were laid under the conditions of a regressive lower Visean sea, whose boundary passed somewhat west of Novo-Pavlovka. The area under study presented a shallow segment of an open sea whose salinity and dynamics were favorable for the development of large brachiopods and solitary corals.

GEOLOGIC HISTORY OF THE AREA IN THE TOURNAISIAN AND LOWER VISEAN

A detailed lithologic study of Tournaisian and lower Visean deposits, developed throughout the area of the northern slope of the Ukrainian crystalline massif, makes it possible to refine somewhat the geologic history of this area, for that time.

Carbonate rocks in the south Donbas districts and along the northern slope of the Ukrainian crystalline massif belong to two sedimentary sequences: the Tournaisian and lower Visean.

The Tournaisian sequence begins with arenaceous and argillaceous deposits which change upward to massive dark gray, microgranular to cryptogranular limestones interbedded with shales of zone C_1^ta. They were deposited in a shallow marine embayment which stretched from the southern Donbas districts in the east to Alefirovka, in the west.

The marine basin gradually expanded, until it reached Novo-Moskovsk, by the time zone C_1^tb was deposited.

Dolomites and dolomitic limestones of zone C_1^tb represent the deposits of a shallowing marine basin of higher salinity. An embayment of that sea, with a somewhat brackish water, existed at that time in the Znamenka and Novo-Moskovsk area.

Tournaisian transgression was at its maximum by the time of deposition of organogenous C_1^tc limestones. The westernmost point reached by the sea, at that time, was Kulebovka Village of the Novo-Moskovsk district.

Zone C_1^td is developed only in the southern Donbas districts where they represent shallow deposits of Turnaisian sea reduced to its minimum.

Deposits of zones C_1^Va-c, which rest on various Tournaisian beds, and on the Precambrian west of Kulebovka, represent the lower series of a greater lower Visean transgression which spread along the northern slope of the Ukrainian crystalline massif, as far as the Podgorodnoye area. The area under study was occupied at that time by a shallow sea of a normal salinity and normal hydrodynamic conditions. Chiefly organogenous calcium carbonate was deposited there, with a periodic and intensive influx of terrigenous argillaceous material. In the western districts, this latter process was so intensive as to create an unfavorable environment for a marine fauna. For this reason, mostly chemical precipitation of $CaCO_3$ took place there.

Siliceous limestones and cherty rocks of zone C_1^Vd-e represent deposits of a gradually contracting and shallowing lower Visean sea, retreating eastward. The westernmost limit of that marine basin was the Pavlograd district.

The lower Visean sea continued to shrink

during the deposition of organogenous clastic limestones of zone C_1^Vf. These deposits spread as far west as the Novo-Pavlovka area. The conditions of an open shallow sea existed in the southern reaches of the Donbas.

In its retreat to the east, the lower Visean sea left behind a rather low marshy coastal plain stretching along the northern slope of the Ukrainian crystalline shield. It was covered with vegetation, the remains of whose root system occur throughout the area, as far as the southern Donbas districts. Peat was formed in many places at that time. Coal beds or rocks with a "curly" structure occur throughout the area; they are overlain by limestone B_1 (A_4 in the Ukrainian Geological Administration nomenclature) which represents the lower member of a new major upper Visean transgression.

There is no evidence of the westward facies replacement of limestone zone C_1^Vf, along the northern slope of the Ukrainian massif, which is supposed to take place there according to D. Ye. Ayzenverg, N. Ye. Brazhnikova, A. Z. Shirokov, and others [1-6]. Regarding the similarity between the faunal assemblages in carbonate rocks of zone C_1^Vf of the southern Donbas districts and those in lower limestone beds of the overlying zone C_1^Vg, in the west, it can be explained by the similarity in the depositional conditions for both. As witness the nature of limestone $B_1(A_4)$, the depositional conditions were similar for limestones in the lower part of the terrigenous-marine sequence, in the west, and for those of zone C_1^Vf, in the east. In the Bogatyr' area and east of there, limestone B_1 is mostly finely clastic, consisting almost entirely of small indeterminate faunal fragments with isolated coarser fragments of brachiopods, crinoids, and foraminifera. Such a limestone was deposited in relatively shallow reaches of the sea, where fine organogenous clastic material was carried from beaches and shoals.

West of there, in the area of Novo-Pavlovka, Alefirovka, Pavlograd, and Novo-Moskovsk, the predominant components of limestone $B_1(A_4)$ are coarse, in places, well-rounded, fragments of brachiopods and crinoid segments. There are isolated foraminifera and algae, in places, with angular to rounded quartz grains, as much as 0.25 mm. Such limestones were deposited in the wave-affected zone of a shallow sea. The accumulation of limestones of zone C_1^Vf, in the southern Donbas districts, proceeded under the same conditions. Accordingly, the similarity in the faunal assemblages from limestones in the lower part of the zone C_1^Vg, in the western districts, and from the C_1^Vf Donbas zone is explained by the similarity in their sedimentary conditions rather than by their allegedly similar age.

At the period of accumulation of the carbonate sequence, the area under study was subjected to fairly slow oscillatory movements on the background of a general submergence. The tectonic conditions of the area changed substantially after the carbonate sequence had been laid down. The accumulation time of zone C_1^Vg was marked by frequent oscillations which determined its cyclic structure.

It is difficult to conceive that tectonic conditions different from the above prevailed at that time in the adjacent segments of the northern slope of the Ukrainian crystalline massif. Most probably, the intensification in the rhythmic oscillation of the earth's crust occurred simultaneously throughout the entire greater Donbas area, including its western districts.

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QUATERNARY DEPOSITS IN THE NORTHERN CASPIAN REGION[1]

by

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In describing the northern Caspian region Quaternary, the author refines the present concept of its stratigraphy. He differentiates the Khvalynsk stage into three formations instead of two. The Singil' and Astrakhan' beds, assigned to the lower Khozarian formation by P. V. Fedorov, are taken out of it. New data are cited on very recent tectonic movements in the area.

* * * * *

The post-Apsheronian deposits of the northern Caspian region have been studied by many geologists, with the most detailed stratigraphic investigation carried out by P. A. Pravoslavley [17-20], M.M. Zhukov [9, 10, and others], V.A. Kovda [14], I.P. Gerasimov, and in recent years by P.V. Fedorov [22], Ye.V. Shantser [23], and others.

V.I. Gromov [8], N.I. Nikolayev, and G.F. Mirchinok studied the mammal fauna from these deposits; P.A. Nikitin, P.I. Dorofeyev, and V.P. Grichuk investigated the plant remains, etc. As a result of this work, stratigraphic columns for the northern Caspian region Quaternary have been compiled, the most important being the correlations by P.A. Pravoslavlev [19, 20], M.M. Zhukov [9], P.V. Fedorov [22], and those approved by the Commission for Quaternary Study, Academy of Sciences, U.S.S.R. [21].

Up to recently, Quaternary study in the northern Caspian region was based on material from natural exposures along the lower Volga and Ural. At the present time, there is voluminous material from numerous boreholes drilled recently between the Volga and Ural.

This paper is an attempt to generalize the data on hand, and to use them together with the author's personal observations, in refining the knowledge of the stratigraphy of Caspian sediments.

* * *

Quaternary deposits are developed everywhere in the north of the near-Caspian depression (within the Volga-Ural watershed) where they are represented by both marine and continental formations. They are very irregular in their lithology and facies and are difficult to correlate even in adjacent sections. The study of a large number of sections revealed a number of features characteristic of different formations, and certain patterns in their vertical change. Because of the lack or scarcity of a marine fauna in these Quaternary deposits, our differentiation of the nearly barren sections in the Volga-Ural watershed is based on such features as the evidence of erosion, the presence of fossil soils, the change in color and composition of the rocks, etc. The age of beds so identified has been established from rare findings of an index fauna and from analogy with Volga, Ural, and Uzen' sections.

BAKU STAGE[2]

Baku deposits are represented by both marine and continental formations. Because they are poorly defined faunally, their dif-

[1]Chetvertichnyye otlozheniya severnogo prikaspiya.

[2]As the basis of our stratigraphic differentiation, we used V. I. Gromov's classification of 1957, where the base of the Quaternary is drawn at the base of the Akchagyl' stage; in that classification the described stages belong to the middle part of the system, the Pleistocene (Q_2). A description of Eopleistocene and Holocene deposits of the northern Caspian region is given in other papers.

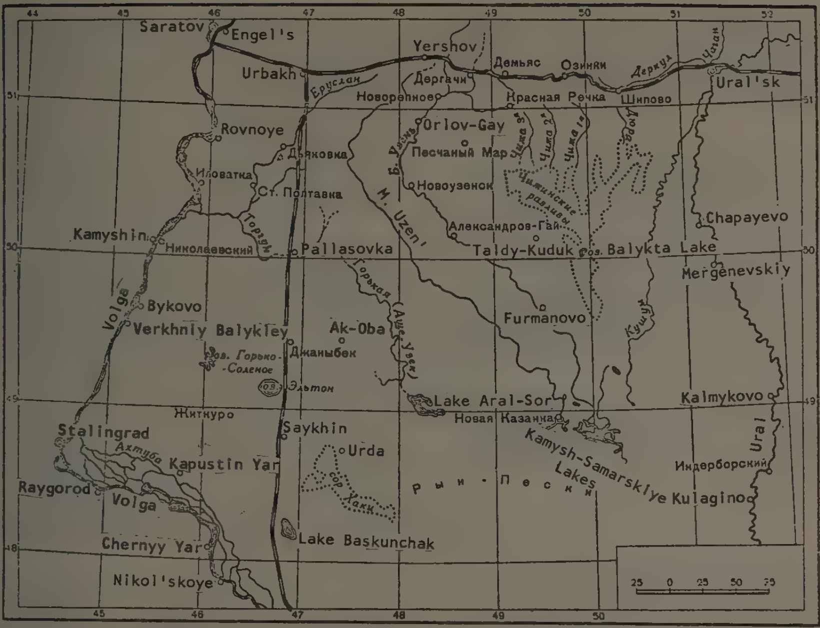

FIGURE 1. Index map of the northern Caspian region.

ferentiation in the area under study is somewhat tentative. Their base is drawn on the top of underlying Apsheron sands, very consistent in their occurrence at a depth of 25-35 m. Their upper boundary is drawn on the top of Astrakhan' loams overlain by dark Khazarian clays. Baku deposits are usually shallow (15-20 m).

Along the lower Volga, they are exposed at Chernyy Yar, where they are represented by dark-gray clays with layers of sands containing Didacna ex gr. catillus Eichw. sp., similar to D. parvula Nal., D. aff. parvula Nal., Adacna plicata Eichw., Dreissena distincta Andrus (idendified by P.V. Fedorov). The clays are flexed into a gentle fold and broken up by a number of normal faults. Along the Volga, however, Baku beds usually lie below the water level, being exposed only in their upper part where they are represented by dark slimy clays (Singil') with shells of Unio sp., Palulina sp., P. vivipara Müll., Valvata fluviatilis Müll., Sphaerium rivicola Leach, and by peatlike to lignitized stems and roots of family Salix [see P.A. Pravoslavlev, 17, p. 28]. P.A. Nikitin found remains of plants Selaginella selaginoides, Menyanthes trifoliata, and Azolla filiculoides, in Singil' beds [15].

According to V.P. Grichuk [7], tsuga pollen and Pinus strobus have been found in Singil' beds at Raygorod. Singil' clays have also been observed at Verkhniy Balykley, absolute elevation 0; at Bykovo at 9 m elevation, and at other places.

East of the Volga, deposits which we assign to the Baku stage were laid down probably on relatively higher ground. They are subdivided here into two layers, with the upper, occurring at 15-20 m, consisting of loams and brown to reddish clays which carry large calcareous inclusions and gypsum. Buried soil has been observed at the top of this bed, 1-6 m thick, in a number of localities (Ak-Oba, Pallasovka Villages). We correlate this part of the section with the Astrakhan' formation of P.A. Pravoslavlev. The lower bed, as thick as 6-8 m, is represented by darker clays with plant remains and shell fragments of fresh-water molluscs which have not been identified. These beds are not everywhere distinct, so that the Baku deposits locally become either more motley or more uniform, depending on sedimentary conditions. In the higher area, apparently inherited from ancient relief (such as near the syrt plain or syrt remnant), light-colored loamy formations are chiefly developed. In depressions they give place to dark clayey beds. It is possible that marine deposits are present in the Elton [11] and in the vicinity of Lake Baskunchak.

Near the Bol'shoy Uzen' River, Baku deposits occur at a depth of 18-20 m, where

they are represented by dark gray to bluish clays. At Furmanovo (Altyn-Bay-Aral), they carry a fauna of Didacna rudis Nal., D. baery Grimm., D. celekenica Andrus., Micromelania and Neritina [4].

In the southern part of the Bol'shoy Uzen' -- Kushum watershed, these beds locally occur at 8-14 m and carry a typical Baku fauna and microfauna. The Baku deposits become thicker, southward.

In the lower course of the Ural, Baku marine beds, represented by sands, are exposed near Lake Inder and at Mergenevskoye Village [10]. At other points, only the continental slimy Singil' clays stand above water.

It appears, then, that marine and generally water-laid deposits are associated with the lowest parts of the area, inherited from the Baku-time relief.

KHAZARIAN STAGE

Khazarian deposits are distributed throughout the area. In the vast majority of instances, they are represented by continental formations and are recognized in the sections by their position between the Baku and Khvalynsk deposits. The top of this stage is drawn on a buried soil underlying the Khvalynsk marine beds; its base is drawn on top of the above-described Astrakhan' beds.

Khazarian deposits occur at a shallow and fairly consistent depth of 4-6 m. They are almost completely exposed along the lower Volga. Downstream from the mouth of the Yeruslan River, Khvalynsk clays are underlain by brownish-yellow loams and grayish clays, 0.5-2m thick, changing downward to light-colored cross-bedded sandy loams and sands, 5-7 m thick. The latter rest with an erosional break upon dark to brownish, slimy clays with a common fresh-water fauna of Unio pictorum Linne, Sphaerium rivicola Leach, and Pisidium amnicum Müll. [18, p. 152 ff]. Thus there are present two main series: the upper sandy loam which we designate as the Atel' or upper Khazarian; and the lower, slimy clay, which we call the Kosozhskoye or lower Khazarian. They are separated by an erosional break.

The same two Khazarian formations occur in the well-known Sukhaya Mechetka exposure (northern suburb of Stalingrad), where the Atel'skiy loams, changing downward to sands, are underlain by a buried marsh-meadow soil which tops the lower part of the Khvalynsk stage. The cultivated layer of this soil yielded Paleolithic artifacts, dating back to late Mousterian, also mammal bones (Canis lupus L., Elephas primigenius Blum,[3] Cervus elephus L., Saiga tatarica L., Bison priscus Boj., etc.)[5]. According to A.I. Moskvitin (oral communication), this soil carries pseudomorphs of ice wedges. The lower Khazarian bed is made up of dark loams, rubbly alluvial sands, correlative with those exposed at Chernyy Yar, as described below. These sands are 3-4 m thick and rest on Paleogene rocks.

At Raygorod Village, the Atel'skiy loams, which change downward to sands, rest with an erosional break on lower Khazarian slimy loams with a fresh-water fauna of Valvata piscinalis Müll., Bithinia sp., and Planorbis marginatus Drap. [17, p. 27]. Developed at the top is a marsh-meadow type soil broken up by (ice) wedges. In its character and position in the section, this soil is correlative with the Mousterian soil in the Sukhaya Mechetka hollow. Lower Khazarian slimy loams change downward to cross-bedded river sands (Chernoyarsk) which rest with a deep erosional break on dark (Singil') clays.

There is evidence of soil making and pseudomorphs of ice wedges, at the top of the Atel'skiy beds, at Chernyy Yar. The Atel'skiy loams rest on an eroded surface of Kosozhskoye (lower Khazarian) deposits, with local remains of a thick marsh-meadow fossil soil. The lower bed is made up of oxbow type sediments: stratified slimy clays with large Unio, as much as 8 m thick, changing along the strike to cross-bedded river sands (Chernoyarsk); the sands contain shell fragments of fresh- to brackish-water molluscs (Didacna of a Khazarian aspect, Dreissena polymorpha Pall., Paludina sp.). The same sands have yielded a whole cranium of Elephas trogonterii (primigenius), also bones of Megaceros germanicus, Bison priscus longicornis, Saiga sp., and Equus (Equus) sp. [8].

The upper part of lower Khazarian beds carries evidence of permafrost. The Chernoyarsk sands rest with an erosional break on dislocated Baku deposits.

Similar sections were observed at Solenoye Zaymishche, Grachi, and Nikol'skoye Villages.

At Vetlyanka Village, the Atel'skiy loams locally are fully replaced by sandstones which in places carry at their base large accumulations of mollusc shells (commonly both valves), probably representing near-shore

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[3]Because of the poor preservation of the bone material, the specific identification of the mammoth is tentative.

Didacna sp., similar to D. pallasi Prav., and D. catillus Eichw., D. pallasi Prav (fragment), Monodacna edentula Pall., Monodaena caspia Eichw., Adacna laeviscula Eichw., A. plicata Eichw., Dreissena polymorpha Pall., Corbicula fluminalis Müll., Sphaerium sp., Pisidium sp. (identified by P.V. Fedorov). These sands are closely related to Atel'skiy loams (which, therefore, should be assigned to the upper Khazarian stage), and are correlative with sands from the lower part of the Atel'skiy, in sections to the north. The sands rest unconformably on eroded stratified greenish loams, with shell fragments of marine lower Khazarian molluscs in the top of the latter.

The generally similar exposure at Kopanovka Village, also contains two beds of Khazarian marine sediments. The upper of the two has been correlated with the Atel'skiy beds, and the lower (Kosozhskoye) has yielded besides the forms cited by V.P. Grichuk [7], Didacna aff. Subovalis Prav., D. ex gr. nalivkini Wass, and others, typical, according to P.V. Fedorov [22] of the upper Khazarian stage.

Similar sections have been observed at Yenotayevka; at Vladimirovka (Yenotayevskaya), according to P.V. Fedorov [22], gray clays wedge out between the Atel'skiy loams and lower Khazarian clays. These gray clays carry an upper Khazarian fauna: Monodacna caspia Eichw. (predominant), Didacna sp. (large flat forms similar to Didacna subovalis Prav. and D. surachanica Andrus.), D. ex gr. crassa Eichw., and rare valves of D. pallasi Prav. Below Vladimirovka, the Khazarian beds dip steeply beneath the water surface.

Thus the two main Khazarian beds are traceable in the lower course of the Volga. Although they both (Kosozhskoye or lower Khazarian, and Atel'skiy or upper Khazarian) carry an almost identical mollusc fauna, including the representatives of both the upper and lower Khazarian beds, we deem it expedient, nevertheless, to separate two stratigraphic units in the Khazarian stage. These differ in their lithology and facies and are separated by a regional erosional hiatus and by a layer of soil on top of the lower unit. The soil, in many places, shows evidence of permafrost.

It is pertinent to note that P.V. Fedorov [22] erroneously took the Kosozhskoye formation of P.A. Pravoslavlev (previously known as the proper Khazarian marine deposits) for the Singil' beds. Because of their position in the section, and their mollusc fauna, he assigned that interval to the lower and partially to the upper layers of the Khazarian stage. In so doing, P.V. Fedorov first of all took all stratigraphic meaning out of the term, Kosozhskoye formation; second, he erred in assigning to the Khazarian stage that part of the section (Singil' and Astrakhan' beds) which is an unalienable portion of the Baku stage.

Along the Ural, below Ural'sk, Khazarian deposits have the same aspect as on the Volga. At Vladimirovka, the Atel'skiy loams change downward to sands of the Chernoyarsk type, which include a lense of lacustrine clays. At Kolovertnoye Village, the Atel'skiy loams are replaced by stratified lacustrine clays and loams. At Mergenevskoye and Kalmykovo, alluvial sands underlying the Atel'skiy loams rest erosionally on dark-blue clays containing plant remains. These clays are correlative with the lower Khazarian or Singil' Volga beds. A marine fauna appears in lower Khazarian deposits only near Lake Inder. Shell deposits in a lense of shell limestone of the kind which occurs at Vetlyanka have been found in alluvial sands at Topoli Village. Below Saraychik, the Khazarian beds dip beneath the water surface.

In the northern Volga-Ural watershed, Khvalynsk deposits are underlain, at a depth of 3-6 m, by yellow-brown loams and sandy loams, with calcareous and gypsiferous inclusions; they are nonstratified, commonly topped by fossil soil. In their lower part, the loams are sandy in many places. In analogy with the Volga sections, we have assigned these deposits to the upper or Artel'skiy bed. In topographic depressions (estuarine lakes, hollows), the yellow-brown loams have been replaced by subaqueous brown to drab-gray loams containing a freshwater fauna. The Atel'skiy loams are underlain by a marsh-meadow fossil soil which is, however, not everywhere present. This soil occurs at a depth of 8-12 m, on top of the lower Khazarian stage which is made up here of drab to brown loams and clays with layers of sandy loams and sands. The rock contains pepperlike inclusions, intricately shaped limy concretions and gypsum nodules. On elevated areas, such as in the vicinity of the foothill terrace, the lower layer is made up of drab-brown loams which are replaced in depressions by dark slimy beds.

This layer, which we correlate with the lower Khazarian from the lower Volga, is 4-12 m thick and is underlain by drab-yellow to drab-red Astrakhan' clays. The Khazarian beds are locally sandy, as for instance at the southern end of the Talas-Kuduk saline (sor), in the south of the right-bank area of the Gor'kaya River, at Lake El'ton, and in the vicinity of Aral-Sor, Zhank-Sor, and Topchak-Sor. In the lower course of the Gor'kaya River, Khazarian deposits carry fragments of gray calcareous sandstones. All these beds

are continental, water deposited, and subaerial.

Marine fauna occurs here only in large topographic depressions, such as at Zhitkur Village, Lake El'ton, Aral-Sor, in the lower course of the Gor'kaya, where the early Khazarian sea penetrated at the maximum transgression. A mixed fauna (marine and fresh water) occurs in the Bersh-Aral basin and in the lower course of the Gor'kaya. This fauna is missing in the Khazarian beds of the adjacent but higher area. Marine lower Khazarian deposits are present along the Bol'shoy Uzen' River where a section similar to that of the Volga has been observed. They are also exposed on the Altyn-Bay-Aral dome and elsewhere.

In the Chizha and other flood plains, the Khazarian stage is represented by drab to motley rocks. Their facies are inconsistent, in both lithology and color, both laterally and vertically. In the area under study, these beds are difficult to separate from stratigraphically higher and lower deposits, let alone differentiating them into beds. Locally the Khazarian deposits carry a marine to mixed fauna associated with dark slimy clays, apparently at the base of the stage. In the Taldy-Kuduk area and south of Shalabay Village, the upper part of the stage becomes appreciably sandy. Khazarian deposits here are 10-16 m thick.

Thus, both along the Volga and within the northern Volga-Ural watershed, Khazarian deposits are thin and more or less definitely divisible into two units consisting of chiefly continental (water-deposited and subaerial) formation. As in the Baku stage, the marine and water-deposited Khazarian sediments are associated with depressions apparently inerited from ancient topography.

KHVALYNSK STAGE

Khvalynsk deposits are exposed nearly everywhere in the area under study. Only in flood plains and estuarine lakes are they buried under thin post-Khvalybsk sediments.

The presence of two marine beds, separated by continental -- the so-called Yenotayevka ("Yenotayevskiye") -- beds, has been established by P. V. Fedorov [22], Ye. V. Shantser [23], Yu. Z. Brotskiy and M. V. Karandeyeva [3]; and earlier by V. A. Kovda [14], I. P. Gerasimov [6], and P. A. Pravoslavlev [18-20].

The following data show that an additional -- middle Khvalynsk -- bed, corresponding to the transgressive Caspian stage, should be separated. This bed is made up of the well-known "chocolate" clays. The individuality of this bed is substantiated by the following conclusions:

1. Along the Volga, Yeruslan, in the El'ton basin, and elsewhere, the middle Khvalynsk chocolate clays form a built-up terrace (standing 20-25 m above sea level), intermediate between the early Khvalynsk northern Caspian Plain (elevation 30-40 m) and the first terrace above the flood plain (upper Khvalynsk) terrace (0-10 m markers). Along the Volga, the middle Khvalynsk terrace is the second above the flood plain.

2. In these areas, the middle Khvalynsk chocolate clays lie adjacent to early Khvalynsk deposits. That was noted by V. A. Kovda [14], M. M. Zhukov [9] and M. P. Britsyna [2]. The juxtaposition is well exposed also in the Lake El'ton basin and at Zhitkur Village. In areas of younger downward movements, chocolate clays are in an erosional contact with upper Khvalynsk deposits.

3. All three marine beds are separated by continental formations. The upper and middle Khvalynsk deposits are separated by the Yenotayevka beds [3, 13, 23]. The middle and lower beds are separated by the El'ton continental beds.

4. The upper surfaces of all three beds do not maintain the same relative position throughout the area, but come together and intersect each other, if traced from stable to sinking areas. This suggests a certain time interval separating the formation of each of these surfaces, under the conditions of persistent negative movements. Thus, where in the Volga valley, near the Yeruslan mouth, the elevation difference for the early and middle Khvalynsk surfaces reaches 10-15 m, and that for the middle and upper Khvalynsk surfaces is 8-10 m, in the south it is 5-7 m and 4-5 m, respectively. Below Chernyy Yar, the chocolate clays no longer adjoin the early Khvalynsk sediments but overlie them, forming topographic highs and the nuclei of Berov's Hills, whereas the upper Khvalynsk sediments here lie on top of chocolate clays.

We believe all this to be sufficient evidence for an individual status of the chocolate clays, as a Khavalynsk bed. In so doing, we concur with the earlier opinion of V. A. Kovda [14] and partially of M. P. Britsyna [2] and N. I. Nikolayev [21] of a younger age of chcoolate clays, as compared with deposits of the maximum lower Khvalynsk transgression.

Lower Khvalynsk bed ($Q_2^{hw_1}$) consists solely of marine deposits whose distribution is confined within the 48 m contour which

passes in the north at the foot of a foresyrt erosional escarpment. At the foot of the latter, Khvalynsk deposits form a near-syrt terrace (elevations 40-48 m above sea level) of a pre-Khvalynsk sea erosion: dustlike drab-yellow, nonstratified, more or less gypsiferous loams, not more than 4-5 m thick.

In the areas of the Volga sandy ridge, Lake El'ton, Lake Bol'shoy Solenyy Sokryl (Great Salt Sokryl), and elsewhere, early Khvalynsk deposits are very sandy, locally forming large sandy massifs. Similar sands are known from Rovnoye Village; and along the Yeruslan River, at D'yakovka.

Lower along the Volga, early Khvalynsk deposits are exposed only in a few places and are usually represented by drab-yellow loams. In the Sukhaya Mechetka ravine, this bed includes sands overlying the Atel'skiy loams and forming a terrace with an elevation of about 30-40 m above sea level. Below Stalingrad, early Khvalynsk loams locally lie at almost the same level with chocolate clays (middle Khvalynsk). Below Chernyy Yar, the upper Khvalynsk bed includes a uniform, although thin layer of sandy loams and sands. The latter lies between the Atel'skiy loams and chocolate clays, separated from either by a more or less distinct erosional break. G. I. Popov [16] believes these sands to be lower Khvalynsk (assigning the maximum transgression to the middle Khvalynsk time) and correlates them with the Karangat bed of the Azov region.

Along the lower Ural, early Khvalynsk deposits occur as they do along the Volga. In the Volga-Ural watershed (with the exception of the Chizha and other flood-plains areas) the lower Khvalynsk stage consists chiefly of drab-gray, light to medium hues, dustlike, gypsiferous and saline loams, of a uniform thickness of 4-6 m.

Middle Khvalynsk deposits ($Q_2^{hv_2}$) are represented mainly by marine to brackish (in the north) deposits of chocolate clays. In the Volga sand ridge area, above Stalingrad, along the lower course of the Yeruslan, these deposits cover the second terrace above the flood plain (elevation 15-25 m), and adjoin the early Khvalynsk deposits. A similar juxtaposition has been observed in the Sukhaya Mechetka ravine, where chocolate clays, more than 10 m thick, fill a large hollow and adjoin the Atel'skiy loams overlain by early Khvalynsk sands. Along the Volga, below Stalingrad, chocolate clays are developed in topographic lows; from Chernyy Yar on, they are developed everywhere, overlying the uniform layer of early Khvalynsk sands.

In exposures at Lenino, Chernyy Yar, Vladimirovka, and other villages, the upper part of the chocolate clay bed exhibits steep folds, the results of permafrost. At Yenotayevka, Vladimirovka, and other places, chocolate clays form the nuclei of Berov's Hills.

In the western part of the Volga-Ural watershed, early Khvalynsk chocolate clays are developed along the Yeruslan, in estuarine lakes of the Tazha, Prishib, Medvezh'ya Rivers, and others, and on the shores of Lake Gor'ko-Solenoye. From that area, the chocolate clays spread out in a narrow terrace along ravines to Lake El'ton and thence toward the vast depression of the Khaki saline. West of the latter, middle Khvalynsk deposits stretch in a wide tongue toward Lake Botkul' and Zhitkur Village, where they also form a middle Khvalynsk terrace.

Middle Khvalynsk deposits are widely developed on the El'ton shores. They dip and thicken toward the lake. In the lake itself, according to G. A. Vasil'yev [4], deposits correlative with the middle Khvalynsk lie at an elevation of 36 m and are 25 m thick.

These rocks are widely developed in the Gor'kaya River basin. Here, terraces made up of them fringe in wide bands the salines of Sorkul', Bersh-Aral, and others. Along the south and east shores of Lake Aral-Sor, chocolate clays have been preserved only in valleys joining the lake, with the middle Khvalynsk terrace missing at its shores. Chocolate clays are also developed in the salines' area, in the Malyy Uzen' -- Gor'kaya watershed. Along the Malyy and Bol'shoy Uzen' Rivers, middle Khvalynsk clays do not build an identifiable terrace but lie almost at the same level with upper Khvalynsk deposits, forming thin and areally restricted lenses in depressions along modern and ancient streams. On the Malyy Uzen', 10 km south of Zhulduz, chocolate clays lie on an eroded drab-yellow loam with Didacna protracta Eichw. (identified by P. V. Fedorov). On the Bol'shoy Uzen', below Aleksandrov-Gay, the erosional pockets between the Atel'skiy loams and chocolate clays carry a thin layer of greenish lumpy clay related to the latter and filled with a fresh-water fauna (El'ton beds). East of the Uzen's, chocolate clays locally occur in estuarine lakes and flood plains.

Along the lower Ural, middle Khvalynsk deposits do not form an individual terrace, either, but lie at the same level as the lower Khvalynsk. The chocolate clays are exposed fairly fully at Gorskoye, Khar'kino, Kalmykovo, etc. At Khar'kino, their thickness increases to 8-10 m. The upper, typically chocolate clays gradually change downward to drab, gray, and finally to bluish-gray (El'ton clays). A similar change has

been observed at Ural'sk.

Upper Khvalynsk deposits ($Q_2^{hv_3}$) are represented on the whole by lacustrine-alluvial and partly marine sediments. The latter are missing in the south and are outlined by the zero to +3m contour. In the lower courses of the Volga, Ural, and Uzen's, they are represented by stratified sandy loams and argillaceous sands. Along the lower Ural, marine upper Khvalynsk deposits have been described by I. P. Gerasimov and A. G. Doskach [6], who have separated as many as three post-Khvalynsk transgressions. Deposits of the oldest Yama -- the Yamanskiy -- are correlative with upper Khvalynsk deposits along the lower Volga. Along the Malyy Uzen', south of Zhulduz, upper Khvalynsk marine beds are represented by sandy loams and sands with Didacna protracta Eichw. and Dreissena polymorpha Pall. (identified by P. V. Fedorov) and form a very broad (as much as 1 km) first terrace.

Outside the area of negative elevations, the upper Khvalynsk bed is represented by alluvial and lacustrine deposits which form the lowest terraces above the flood plain (Kushum, Sarpi) of streams and are also developed in estuarine lakes and flood plains.

The upper Khvalynsk stage appears to include the dark-colored rocks locally developed along the two Uzen's. Along the Bol'shoy Uzen', from Berezin to Port-Arthur, and along the Malyy Uzen' from Lokhmatovka to Kok-Terek, dark to black, gley loams and sandy loams, locally greenish gray, ocherous, with fresh-water mollusc shells, rest above marine Khvalynsk beds, with a slight erosional contact, as seen in the upper parts of river cliffs. Their thickness is 1-3 m.

Sediments of the same type but not as well represented occur along the Volga, at Bykovo and Chernyy Yar. They are developed only in topographic lows and in estuarine lakes and deeps. Being older than the adjacent alluvium of the first terrace above flood plain, they appear to be water-deposited equivalents of the subaerial Yenotayevka beds, having been formed during the middle Khvalynsk regression, prior to the pre-late Khvalynsk erosion.

* * *

The above data makes possible the construction of a stratigraphic outline (Fig. 2) which is very similar to that of P. A. Pravoslavlev [20] but differs somewhat from the present stratigraphic outlines for the northern Caspian-region Quaternary.

Likewise, the new data suggest certain conclusions as to the character of the deposits and of the most recent movements of this area.

1. Quaternary deposits of the Volga-Ural watershed differ greatly from those in the lower courses of these rivers. Thus alluvial Chernoyarsk sands are missing in the watershed where they are replaced by subaerial loams. Lower Khazarian and partly Baku water-deposited sediments, developed along the Volga and Urals, are replaced in the watershed by subaerial deposits different in their aspect. The same is true for middle and late Khvalynsk deposits. Such a difference in the sections makes their correlation very difficult. This is the reason for gross errors in the correlation of the Volga and the watershed sections on purely lithologic evidence. For instance, a number of authors designate as Khazarian sands the thick (more than 30 m) sandy sequence occurring at a depth of 25-30 m; they also assign to the Khvalynsk stage all of the upper loam-clay mantle of the watershed, 20-25 m thick; and designate as Baku the clays occurring at a depth of more than 100 m.

2. Because the index Quaternary fauna is comparatively rare in the Volga-Ural watershed deposits, great stratigraphic significance is ascribed to such details as fossil soils; the change in coloring, lithology, and the character of concretions; to erosional surfaces, etc. All reflect regional changes in the physical geographical and facies conditions of sedimentation. The index Quaternary beds of the Volga-Ural watershed are drab-red to drab-yellow Astrakhan' loams, the soil and erosion at the base and top of lower Khazarian deposits, and the soil at the top of Atel'skiy loams.

3. In correlating the Quaternary deposits of this area, it should be kept in mind that they, regardless of their age, have undergone radical facies changes in passing from topographic highs to lows. Where in the highs they represented chiefly by subaerial deposition, they are fresh water to marine, in the lows.

4. The various degrees of increase in sand content for various beds have been described above. This increase is of a local character, occurring most commonly in the vicinity of salt domes where older sandy beds have been exposed by the younger uplifts. Thus, a local increase of sand content in sediments may be used as indirect evidence of buried salt domes.

5. Both the thickness and the depth of Quaternary deposits in the northern Caspian region are not great. Thus Baku deposits occur at a depth of only 20-30 m, and even at 10-15 m in a number of localities; they

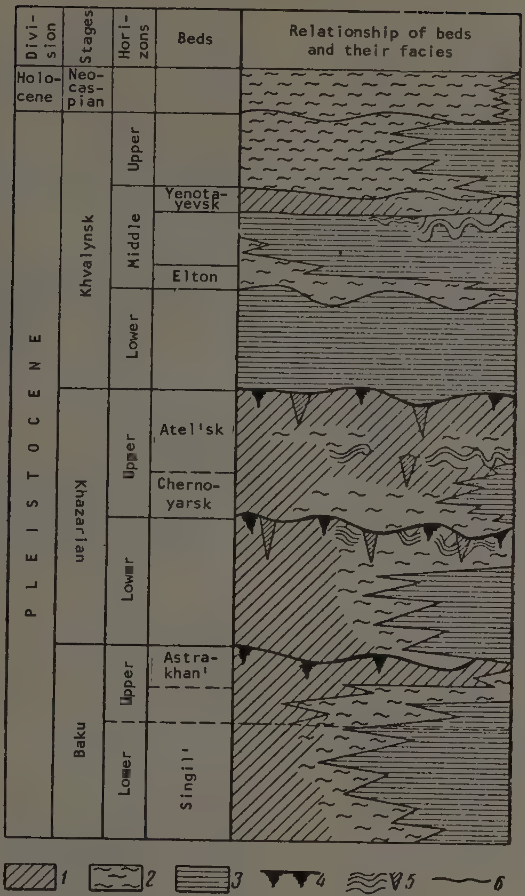

FIGURE 2. Stratigraphic outline for Quaternary deposits in the northern Caspian region.

1 -- subaerial deposits; 2 -- fresh-water deposits; 3 -- marine deposits; 4 -- evidence of soil formation; 5 -- permafrost deformation of beds; 6 -- erosional surface.

are about 8-15 m thick. Khazarian deposits occur at 3-5 m and are 12-18 m thick. Up to now, the prevailing and quite groundless opinion has been that the Baku sediments are here several tens of meters thick and occur at a depth of about 100 m and more. The thickness of Khazarian deposits has been determined as 30-70 m, with as much as 30 m for the Khvalynsk. Hence the erroneous conclusions on the magnitude of recent sinking in the area: a portion of upper Pliocene deposits was erroneously assigned to various Quaternary stages. The error of these old concepts is clearly seen from the shallow occurrence of faunally characterized Apsheronian, Baku, and Khazarian beds in tectonically undisturbed areas. Thus, Apsheronian deposits are either exposed or occur at depths of 15-25 m, at Novouzensk, Mergenevskiy Village, in the Kamysh-Samara flood plains, in the lower course of the Gor'kaya River, in the basin of both Uzen's, etc. The upper part of the Baku stage is exposed everywhere along the Volga and Ural Rivers, and the two Uzen's. Seen in this light, the data of V.G. Kamysheva-Yelpat'yevskaya [12] are of interest. She mentions the remains of an Apsheronian and Baku microfauna in beds

tentatively assigned to the Khazarian and Khvalynsk stages.

6. The thickness of Quaternary deposits in the northern Caspian region is usually small (15-18 m). Its changes, insignificant on the whole, are determined mostly by local features. For this reason, no conclusions as to neotectonism can be made here, solely from the thickness change. This is even more true for the changes in thickness of individual stages, as formerly understood.

7. Quaternary deposits have been tectonically disturbed over salt domes. Thus, the Baku deposits are intensively disturbed in the vicinity of Lake Baskunchak, at Chernyy Yar. The Khazarian deposits have been brought to the surface along the shores of Lakes El'ton, Baskunchak, and Aral-Sor. A younger sinking in these areas is suggested by the character of occurrence for middle Khvalynsk chocolate clays. It appears, then, that tectonic movements persisted during the Pleistocene in the salt domes, with the greatest intensity manifested in pre-Khazarian and Khvalynsk time.

8. The change in the character of Quaternary deposits and in their elevation, along with the present relief features of the area, suggests the continuity of regional Pleistocene movements as well, throughout the area. Assuming the western part of the Volga-Ural watershed (between the Volga and Ural Rivers) as comparatively stable on the whole, we must ascribe some uplift to the syrt remnant area, to the Volga sandy ridge, and to the Murat-Say and Bersh-Aral ravines watershed, inasmuch as these areas of a relatively high relief consist chiefly of subaerial deposits. On the other hand, the areas along the lower Yeruslan, Lake El'ton, the Lake Botkul' depression, and the Khaki saline, with water-deposited sediments including marine, have been subject to sinking. Areas of the Chizhinskiy, Dyurinskiy and Balykta flood plains, the lower course of the Gor'kaya River, etc., where water-deposited, chiefly marine, sediments were deposited during the Quaternary, are the sites of gradual submergences.

On the background of these general oscillatory movements, there appear both positive and negative local movements affecting comparatively small areas. These minor movements, in their turn, are complicated by the growth of salt domes and by their compensating rim synclines. It can be stated that from Baku time and to the present, the downwarping focus for the eastern part of the Volga-Ural watershed has been the area of the Chizhinskiy and Balykta flood plains, as witness the lack of development of marine deposits in the adjacent area.

Not to be overlooked is the steady and consistent southeasterly tilting of the area, throughout the Pleistocene. This is the direction of the dip and of the thickening of all beds and stages, and of the plunge of the corresponding depositional surfaces. The magnitude of the dip is in direct proportion to the age of the deposits. In addition, the replacement of continental Pleistocene deposits by the marine proceeds in the same direction.

9. Briefly, the general outline of Pleistocene paleogeography is as follows. In the Baku age, the sea advanced considerably to the north, forming wide transgressive embayments in pre-Baku topographic lows. The coast line passed somewhat to the north of the line: Kapustin Yar -- El'ton -- lower Gor'kaya -- Furmanovo -- Mergenevskiy. In depressed segments of the land, water-deposited Singil' clays were formed at that time. During the regression, they were laid over the marine Baku sediments. Chiefly subaerial yellow- to reddish-brown clays and loams were formed on elevated segments of the land. Subaerial rocks were especially widespread at the close of the Baku time, when they formed the Astrakhan' bed.

The onset of the Khazarian witnessed the intensification of erosional processes. Then the early Khazarian sea invaded the topographic lows, along which it encroached deep into the land. The contemporaneous shore passed through sites of Chernyy Yar -- Saykhin Station -- Lake El'ton -- Khaki saline -- local Gor'kaya River -- Aleksandrov-Gay -- Chinzhinskiy flood plains. Yellow-to drab-brown loams of the syrt-type argillaceous deposits were laid down on elevated segments, with fresh-water slimy loams in the depressions. Soil of a marsh-meadow type was developed almost everywhere in the area, by the close of early Khazarian time. Only the Chizhinskiy flood area, the most depressed, witnessed a continuous deposition of water-laid sediments. Then occurred another intensification of erosional processes followed by the accumulation of alluvial deposits in river valleys. Lacustrine and marsh deposits were laid down in topographic lows of the watershed, and subaerial loams and sandy loams formed on elevated areas. At the same time, there was a comparatively minor upper Khazarian transgression in the lower part of the area. It did not advance beyond Nikol'skoye, along the Volga, and reached as far as Topoli, along the Ural. By the close of late Khazarian time, subaerial loesslike Atel'skiy loams were accumulating everywhere, except for depressions where they were replaced by marsh or lacustrine sediments. A fossil soil is developed on top of upper Khazarian deposits.

At the onset of Khvalynsk time, a brief

intensification of erosional processes was followed by the early Khvalynsk transgression, when the sea rose up to the present +38 m mark. As a result of intensive erosion during the following regression, extensive shallow ravines were formed throughout the area. In the beginning of middle Khvalynsk time, they were filled by water-deposited slimy, commonly saline, formations with a fresh-water fauna (El'ton beds). These continental deposits change upward, without a break, to chocolate clays -- marine to brackish deposits of the middle Khvalynsk basin. The latter formed long inlets which were the accumulation sites for the chocolate clays. Such inlets were associated with valleys of the Volga, Ural, Gor'kaya Rivers, the two Uzen's and with depression of the present lakes, salines and estuarine lakes. Continental Yenotayevka beds were formed at the close of the middle Khvalynsk. The expansion of the late Khvalynsk sea was limited by the present zero contour, with the formation of restricted embayments along river valleys. In post-Khvalynsk time, the sedimentation throughout the area was confined to flood plains, estuarine lakes, and river valleys, where lacustrine and alluvial sediments were deposited. The neo-Caspian transgression did not encroach upon the area.

10. The presence of beds carrying evidence of permafrost; the well-known findings of paleolithic artifacts, dating back to the Mousterian (Sukhaya Mechetka ravine); the findings of fauna in the Quaternary of the lower Volga; and the general character of the above-described deposits, make it possible to correlate the post-Apsheronian history of the northern Caspian region with the glaciation in the Russian plain (as set forth in the A. I. Moskvitin chronology, 1957).

We propose the following general correlation: lower Khavarian deposits were accumulated chiefly during the Likhvinskoye interglacial period. The onset of the late Khazarian coincided with an intensive cooling, and with the formation of numerous and pronounced cryoturbations at the top of the lower Khazarian. This cooling is correlative with the Dnieper glaciation, which is confirmed by the findings of Mousterian artifacts, at the top of lower Khazarian deposits. The soil layers at the base of Atel'skiy beds were formed during the Odintsov age. The next cooling, apparently correlative with the Moscow glaciation, brought about the permafrost disturbances in the middle interval of the Atel'sk beds (Nikol'skoye, Kopanovka). The Mikulinskoy interglacial period witnessed the soil formation at the top of the upper Khazarian (Atel'skiy) bed with the (ice) wedges breaking up this soil formed in the Kalinin age. The permafrost mashing of clays in the upper part of the middle Khvalynsk layer, and some of the accumulation of Yenotayevka subaerial loams took place at low temperatures, probably brought about by the Ostashkovo glaciation.

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HYDROTHERMAL METASOMATISM OF PROTEROZOIC ROCKS IN THE OLEKMA-VITIM HIGHLANDS[1]

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This paper describes the metasomatically altered Proterozoic rocks in the Olekma-Vitim highlands, affected by albitization, biotitization, and K-feldspathization. These processes took place in sedimentary rocks previously subjected to contact metamorphism. In a number of instances, the character of alteration depended on the lithologic composition of the original rocks. The metasomatism was accompanied by an extensive migration of the components; a relationship is noted between the addition of individual components, in some rocks, and their leaching from others.

* * * * *

Proterozoic metasedimentary rocks of the Olekma-Vitim highlands underwent intensive metasomatic transformations in some regions. Besides alterations by granitization and normal contact metasomatism, processes of hydrothermal, chiefly alkaline, metasomatism were very active. This paper deals with the description of these latter processes.

The phenomena of hydrothermal metasomatism took place in rocks of the Udokan series, widespread in the Olekma-Vitim highlands where it fringes the Aldan shield in the south and west. The rocks of this series are developed in the Udokan, Kalar, and Kodar Ranges; they are presumably of Proterozoic age, about 10,000 m thick, and varied in their lithology.

As established for the Udokan and Kalar Ranges, formations at the base of the series are made up chiefly of metashales with sandstone intercalations, and to a smaller extent of crystalline limestones. Rocks in the middle part of the section are flyschlike, being represented chiefly by siltstones and sandstones, commonly calcareous and in a few places ferruginous (with magnetite); metashales are uncommon. The upper formations consist of numerous sandstones, including the ferruginous (magnetite), with dolomites and marblelike limestones much more common than in the lower beds.

Most of the rocks have undergone a comparatively slight regional metamorphism. As a result, there are varieties characteristic of the green schist facies. In the Kodar Range, however, as well as in isolated segments of the Udokan Range, some of the formations -- especially those at the base of the series -- were subject to a more intensive and ultrametamorphism resulting in the formation of mica-garnet schists, paragneisses, and migmatites [6].

The Udokan series is cut by intrusions of various compositions, belonging to several phases of a Precambrian magmatic cycle. The oldest are small intrusions of gabbro-norites and diorites. We have assigned to a later magmatic complex the Kodar granitoids in many major intrusions. Minor intrusions of alkali syenites occur in isolated area. In analogy with the Aldan shield alkali rocks, we have assigned them to Mesozoic igneous rocks. Biotite hornfelses, less commonly the biotite-cordierite and biotite-andalusite, are developed at intrusive contacts.

In intrusive exocontacts, rocks of the upper and middle parts of the Udokan series have locally undergone intensive metasomatic alterations such as albitization, quartzitization, biotitization, and K-feldspathization. Both the metasomatically altered rocks and the intrusions themselves are in places associated with major tectonic zones. Within these regional structures, metasomatic rocks are definitely related to minor disturbances, such as zones of shattering and fracturing, originating in the rejuvenation of older faults.

The intensity of these processes is different in different areas.

[1]Gidrotermal'nyy metasomatoz proterozoyskikh porod olekmo-vitimskoy gornoy strany.

The Udokan Range witnesses mostly the processes of albitization, quartzitization, and biotitization, and to a considerably smaller extent, of microclinization. Here, the processes of hydrothermal metasomatism involved rocks of the middle and upper parts of the Udokan series, represented by sandstones, stiltstones, and metashales, with calcareous facies among them. These rocks form anticlinal and synclinal folds, trending northeast, usually with steep dips of 50 to 55°. The folded structures are complicated by faulting. The major faults are almost latitudinal, transversal to the folds. Minor bedding faults are frequent.

The metasedimentary rocks are cut by major intrusions of the Kodar granitoids and by minor intrusions of alkali syenites. Albitization and quartzitization, as well as the associated biotitization, usually occur at the exocontact of granitoid intrusions, either within the hornfels segments or at a considerable distance away from them. These changes affect the rocks in anticlinal limbs, and are in many places associated with small faults. Metasomatically altered rocks commonly show evidence of a later shattering. Microclinization of the enclosing rocks occurs chiefly at the immediate contact with granitoids or alkali syenites.

A description of the several varieties of metasomatically altered rocks of the region is given below.

ALBITIZED AND QUARTZITIZED ROCKS

Both siltstones and sandstones, including the ferruginous, have been affected by albitization and quartzitization. Intensive albitization is common along minor faults, where it leads to the formation of albitities. The latter make up small tabular to tabular bodies, irregular in form, usually nearly conformable with the enclosing slightly metamorphosed calcareous siltstones, at whose expense they have commonly been developed.

The siltstones are dark gray, lepidogranoblastic, consisting of quartz (35 to 40%), acid plagioclase (30 to 35%); biotite, chlorite, and sericite (combined total, 25 to 30%), and ore minerals (about 3%). Some varieties contain carbonate (5 to 10%).

The albitites are pink to bright-red rocks consisting of albite (60 to 70%), quartz (15 to 20%), and varied amounts of carbonates, chlorite, and hematite. Apatite and rutile are present as impurities. The rocks are granoblastic, coarser grained than the original siltstones. Albite (No. 5) forms prismatic to isometric crystals, 0.2 to 0.7 mm across. Quartz occurs either in fine isometric grains dispersed among the albite crystals, or in large segments with numerous growths of small euhedral albite grains. Carbonate idioblasts, as much as 10 mm across are observed in isolated segments (Fig. 1). The carbonate is represented by ankerite (α about 1.530; β about 1.698).

Chlorite is usually formed at the expense of ankerite, first in the peripheral part of crystals, then completely replacing the latter (Fig. 2).

Hematite usually permeates the albite grains, in very fine dust, resulting in the red color of albitites. In addition, it occurs in accumulations of thin scales, locally completely replacing larger carbonate crystals.

FIGURE 1. Albitite with ankerite porphyroblasts. One-fifth natural size.

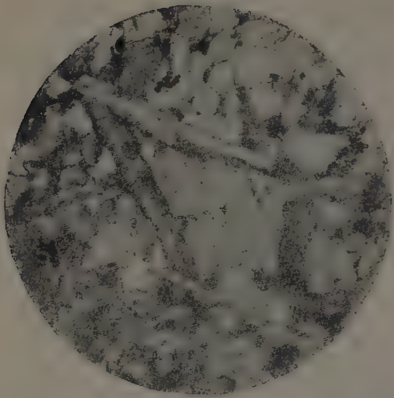

FIGURE 2. Albite with carbonate porphyroblasts. 15X; crossed Nicols.

To give an idea of the addition and leaching of individual components in the albitization process, we have computed the amount of each oxide, in grams to 100 cm^3, for the original and the altered rocks. For this purpose, the percent content by weight, as obtained from the chemical analysis and reduced to 100, has been multiplied by the volume weight of the rock. The last columns (of Table 1) give the increase (+) or decrease (-), in grams, in the amount of each oxide in the transformation of 100 cm^3 of one rock into the other. The algebraic sum of figures in these columns gives the difference in volume weight for the rocks in question.[2]

It appears from Table 1 that a slight albitization and quartzitization of siltstones is accompanied by a slight decrease in their SiO_2 and Na_2O content, by some decrease in the Al_2O_3 content; and by leaching of FeO, Fe_2O_3, CaO, MgO, and K_2O. A more intensive albitization brings about a considerable increase in the Na_2O and in some places in Al_2O_3 content, a decrease in the SiO_2 content, and in a few places a considerable leaching of Fe_2O_3, FeO, MgO, and K_2O.

Ferruginous sandstones undergo great changes in quartzitization and albitization. They are dark-gray dense rocks usually carrying thin (from a few mm to a few cm) intercalations of magnetite, tourmaline,

[2]This order of computation is maintained for the other rocks.

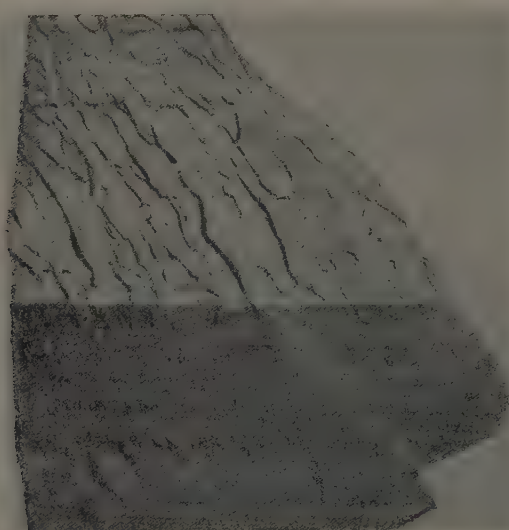

FIGURE 3. Albitized ferruginous sandstone with redeposited biotite and new formations of pyrite. Four-fifths natural size.

Table 1

The balance of matter in the albitization of siltstones

Oxides	% by weight				% by weight reduced to 100				Weight of oxide in grams for 100 cm^3 of rock				Change in oxide content, in g, in the transformation		
	I	II	III	IV	I-a	II-a	III-a	IV-a	I-b	II-b	III-b	IV-b	I to II	I to III	I to IV
SiO_2	68,50	74,56	65,59	62,16	68,60	74,95	65,82	62,19	183,85	194,12	161,92	148,01	+10,27	—21,93	—35,84
TiO_2	0,30	0,22	0,36	0,30	0,30	0,22	0,36	0,30	0,80	0,57	0,89	0,72	—0,23	+0,09	—0,08
Al_2O_3	16,88	14,45	22,96	18,87	16,91	14,53	23,04	18,88	45,32	37,63	56,68	44,93	—7,69	+11,36	—0,39
Fe_2O_3	1,52	0,77	not found	0,96	1,52	0,78	—	0,96	4,07	2,02	—	2,29	—2,05	—4,07	—1,78
FeO	2,14	1,37	0,08	0,43	2,14	1,38	0,08	0,43	5,73	3,58	0,20	1,02	—2,15	—5,53	—4,71
MnO	—	traces	—	0,02	—	traces	—	0,02	—	—	—	0,05	—	—	+0,05
CaO	1,14	0,74	0,70	3,12	1,14	0,75	0,70	3,12	3,06	1,94	1,72	7,43	—1,12	—1,34	+4,37
MgO	2,18	1,40	0,22	1,08	2,18	1,41	0,22	1,08	5,84	3,65	0,54	2,57	—2,19	—5,30	—3,27
Na_2O	4,00	4,35	7,46	8,18	4,01	4,37	7,50	8,19	10,75	11,32	18,45	19,49	+0,57	+7,70	+8,74
K_2O	1,70	0,65	0,90	1,03	1,70	0,65	0,90	1,03	4,56	1,68	2,21	2,45	—2,88	—2,35	—2,11
H_2O	0,10	0,20	0,92	0,19	0,10	0,20	0,92	0,19	0,27	0,52	2,26	0,45	+0,25	+1,99	+0,18
Loss in heating	1,40	0,75	0,46	0,22	1,40	0,76	0,46	0,22	3,75	1,97	1,13	0,52	—1,78	—2,62	—3,23
CO_2	not det.	not det.	not det.	3,39	—	—	—	3,39	—	—	—	8,07	—	—	+8,07
Total	99,86	99,46	99,65	99,95	100,00	100,00	100,00	100,00	268,00	259,00	246,00	238,00	—9,00	—22,00	—30,00
Volume wt.	2,68	2,59	2,46	2,38											

I -- siltstone, dark gray (M. M. Stukalova, analyst); II -- quartzitized and slightly albitized siltstone, light gray with a rosy tint; in contact with I; III -- albitite, rosy red (O. P. Boyarshinova, analyst); IV -- albitite, red with carbonate porphyroblasts (V. M. Kovyazina, analyst).

NOTE: Comma represents decimal point.

Table 2

The balance of matter in quartzitation and albitization of ferruginous sandstones

Oxides	% by weight				% by weight reduced to 100				Weight of oxide in grams for 100 cm³ of rock				Change in oxide content, in g, in the transformation	
	V	VI	VII	VIII	V-a	VI-a	VII-a	VIII-a	V-b	VI-b	VII-b	VIII-b	V to VI	VII to VIII
SiO_2	59,42	66,80	56,06	64,08	59,05	66,64	56,14	64,26	156,48	169,93	149,33	168,36	+13,45	+19,03
TiO_2	0,22	0,16	0,80	0,50	0,22	0,16	0,80	0,50	0,58	0,41	2,12	1,31	—0,17	—0,81
Al_2O_3	20,57	19,67	20,08	10,80	20,44	19,62	20,11	10,83	54,17	50,03	53,49	28,38	—4,14	—25,11
Fe_2O_3	0,47	0,36	5,14	3,38	0,47	0,36	5,15	3,39	1,24	0,92	13,72	8,88	—0,32	—4,84
FeO	1,85	0,35	1,66	1,38	1,84	0,35	1,66	1,38	4,88	0,89	4,42	3,62	—3,99	—0.80
MnO	0,06	0,03	0,02	0,16	0,06	0,03	0,02	0,16	0,16	0,08	0,05	0,42	—0 08	+0,37
CaO	1,00	1,33	1,18	5,10	1,00	1,33	1,18	5,12	2,65	3,39	3,13	13,41	+0,74	+10,28
MgO	5,20	0,68	4,78	4,69	5,17	0,68	4,79	4,71	13,70	1,73	12,72	12,34	—11,97	—0.38
Na_2O	6,70	8,48	2,75	2,40	6,66	8,46	2,75	2,41	17,65	21,57	7,32	6,32	+3,92	—1,00
K_2O	1,00	1,58	4,86	1,70	1,00	1,58	4,87	1.70	2,65	4 03	12,96	4,45	+1,38	—8,51
H_2O	0,08	0,09	0,23	0,26	0,08	0,09	0,23	0,26	0,21	0,23	0.61	0,68	+0,02	+0,07
Loss in heating	3,72	0,70	2,30	5,26	3,70	0,70	2,30	5,28	9,81	1.79	6,13	13,83	—8,02	+7,70
BaO	0,31	—	—	—	0,31	—	—	—	0,82	—	—	—	—0,82	—
Total	100,60	100,23	99,86	99,71	100,00	100,00	100,00	100,00	265,00	255,00	266,00	262,00	—10,00	—4,00
Volume weight	2,65	2,55	2,66	2,62										

V -- slightly albitized magnetite sandstone with preserved dark magnetite bands (M. M. Stukalova, analyst); VI -- more intensively quartzitized and albitized segment of the same rock, with the banding gone (M. M. Stukalova, analyst); VII -- banded magnetite sandstone (N. V. Lodochnikova, analyst); VII -- same, from a quartzitized segment (N. V. Lodochnikova, analyst).

NOTE: Comma represents decimal point.

apatite, in a few places of rutile and ilmenite. Intervals between these intercalations consist of fine-grained aggregates of quartz with a varied amount of carbonates, biotite, and sericite.

Both the quartzitization and albitization considerably lighten the color of the rock and cause the disappearance of its banding. First, the iron oxides are leached out, with some of the magnetite redeposited in isolated grains, immediately outside the magnetite bands. Biotite either completely disappears or else is redeposited in cleavage fractures which cut the rock at a sharp angle to the bedding (Fig. 3). Newly formed pyrite formations are locally associated with these biotite accumulations. Only the banded accumulations of rutile are left from the original magnetite intercalations. In the more intensive alteration and recrystallization, rutile, too, is distributed more evenly throughout the rock.

Rocks with a low magnetite content also were analyzed (Table 2). Their alteration is accompanied by an increase in SiO_2 and locally in Na_2O, with a varied intensity of leaching for Al_2O_3, Fe_2O_3, FeO, MgO, and in places, K_2O. The increase in CaO and heat losses are brought about in places by carbonization of the rock.

BIOTITIZATION OF THE ROCK

Biotitized rocks are comparatively numerous along with the albitized rocks in regions of metasomatism. In places, dolomitic limestones, in thin intercalations and lenses in sandstones and siltstones, are intensively biotitized. In this process, light-gray dolomitic limestones in peripheral parts of the lenses and at contacts with the enclosing siltstones and shales are altered to dark-gray dense rocks. Pres- among them are small carbonate lenses, fully replaced with biotite. Ferruginous biotite ($\gamma = 1.643$), which is developed in the dolomite, forms isometric scales, 0.05 to 0.1 mm; in segments of shattered rocks, it forms veins of coarser (2-3 mm) scales.

A chemical analysis (Table 3) discloses the addition of K_2O, Al_2O_3, Fe_2O_3, and SiO_2, accompanying the formation of biotite. The amount of Fe_2O_3 possibly increases in the oxidation of FeO; CaO and CO_2 are partially leached out.

Mottled sericite and biotite schists are widely developed in isolated segments near the contact with granitoid intrusions. Newly formed biotite occurs as rounded spots, 1 to 3 mm (Fig. 4). These spots also occur in forms close to hexagonal and rhombic. Locally, oval inclusions of finely tabular biotite have a reaction fringe of fine-grained quartz and K-feldspar (nonlatticed). In other instances, these inclusions exhibit a zonal structure, the result of an alternation of biotite and quartz-feldspar bands. It is not uncommon that biotite within a spot forms only a narrow peripheral fringe, whereas the central part consists of one or more comparatively large microcline grains. There are many instances of well-defined biotite acccumulations formed at the expense of carbonate porphyroblasts. Grains of pyrite and chalcopyrite are common in the central part of biotite accumulations.

Not uncommon in ferruginous sandstones is the intensive development of biotite in intercalations enriched with magnetite; this

FIGURE 4. Mottled biotite schists.
Four-fifths of natural size.

Table 3

The balance of matter in biotitization of dolomitic limestones

Oxides	% by weight		% by weight reduced to 100		Weight of oxide in grams for 100 cm^3 of rock		Change in oxide content, in g, in the transformation
	IX	X	IX-a	X-a	IX-b*	X-b	IX to X
SiO_2	15,82	39,96	15,87	39,79	40,15	109,02	+68,87
TiO_2	0,14	0,11	0,14	0,11	0,35	0,30	—0,05
Al_2O_3	4,96	10,78	4,97	10,74	12,57	29,43	+16,86
Fe_2O_3	trace	2,53	trace	2,52	—	6,90	+6,90
FeO	2,95	1,69	2,96	1,68	7,49	4,60	—2,89
MnO	0,73	0,47	0,73	0,47	1,85	1,29	—0,56
CaO	23,92	15,85	23,99	15,78	60,69	43,24	—17,45
MgO	14,61	10,66	14,67	10,62	37,12	29,10	—8,02
Na_2O	1,17	1,39	1,17	1,39	2,96	3,81	+0,85
K_2O	0,48	3,25	0,48	3,24	1,22	8,88	+7,66
H_2O	0,10	0,27	0,10	0,27	0,25	0,74	+0,49
Loss in heating	34,81	13,45	34,92	13,39	88,35	36,69	—51,66
Total	99,69	100,41	100,00	100,00	253,00	274,00	+21,00
Volume weight	2,53	2,74					

IX -- unaltered dolomitic limestone from a nodule (O. P. Baklanova, analyst); X -- biotized dolomitic limestone (V. M. Kovyazina, analyst).

* Column head printed as X-b in Russian text.

NOTE: Comma represents decimal point.

development is accompanied locally by albitization of the intervening intervals.

A chemical analysis of biotite schists (Table 4) shows a lower SiO_2 and Na_2O content, as compared with other metasomatically altered rocks, and a higher content of FeO, Fe_2O_3, MgO, and K_2O.

In contrast with the albitized rocks, K-metasomatism is intensively developed here.

MICROCLINIZED ROCKS

The exocontact of one of the granitoid massifs exhibits intensive microclinization of ferruginous sandstones, with thin intercalations of magnetite maintaining their original position. The remaining rock has been replaced by granoblastic microcline.

Microclinization of siltstones has been observed at the contact with augite syenites. Here, medium-grained microcline rocks with a small addition of quartz have formed among fine-grained hornfelsitic siltstones. It is obvious that the microclinization affects rocks which have been altered by contact metamorphism. Microclinization is especially intensive in the vicinity of pegmatite veins with which this process is probably connected.

In the Kalar Range area, hydrothermally metasomatized rocks of the Udokan series also are represented by metashales and their calcareous varieties. Less common are layers of altered limestones. All these rocks occur in folds trending northeast with dips of 35 to 55°.

The metasomatic alteration of metasedimentary rocks proceeded chiefly in the exocontact of the Chineyskaya gabbroid intrusion. This small intrusion is somewhat elongated along the trend of major folds. Its central part is associated with an anticlinal fold

Table 4

Chemical composition of biotitized metashales and siltstones

Oxides	% by weight			% by weight reduced to 100			Weight of oxide in grams for 100 cm^3 of rock		
	XI	XII	XIII	XI-a	XII-a	XIII-a	XI-b	XII-b	XIII-b
SiO_2	56,30	55,88	57,93	56,28	55,88	57,66	153,64	156,46	159,72
TiO_2	0,88	0,60	0,52	0,88	0,60	0,52	2,40	1,68	1,44
Al_2O_3	18,68	17,93	22,22	18,67	17,93	22,12	50,97	50,20	61,27
Fe_2O_3	6,40	4,44	2,44	6,39	4,44	2,43	17,45	12,43	6,73
FeO	1,48	3,37	2,05	1,48	3,37	2,04	4,04	9,44	5,65
MnO	None	0,05	None	None	0,05	None	—	0,14	—
CaO	1,84	1,65	0,43	1,84	1,65	0,43	5,02	4,62	1,19
MgO	5,25	5,89	3,64	5,25	5,89	3,63	14,33	16,49	10,06
Na_2O	2,70	0,33	0,94	2,70	0,33	0,94	7,37	0,92	2,60
K_2O	3,90	6,48	7,44	3,90	6,48	7,41	10,65	18,15	20,53
H_2O	0,42	0,22	0,10	0,42	0,22	0,10	1,15	0,62	0,28
Loss in heating	2,19	2,84	2,73	2,19	2,84	2,72	5,98	7,95	7,53
SO_3		0,21			0,21			0,59	
P_2O_5		0,11			0,11			0,31	
Total	100,04	100,00	100,44	100,00	100,00	100,00	273,00	280,00	277,00
Volume weight	2,73	2,80	2,77						

XI -- sericite metashale with biotite accumulations fringed with fine microcline grains (Ye. G. Alyaskerova, analyst); XII -- biotitized metashale, nonmottled, homogeneous (M. M. Stukalova, analyst); XIII -- biotitized siltstone (M. M. Stukalova, analyst).

NOTE: Comma represents decimal point.

complicated by faulting.

In a tectonically weakened zone trending northeast, outside the hornfels fringe of the Chineyskaya intrusion, individual beds of sandstones, and metashales are intensively albitized and amphibolized. The trend of this weakened zone coincides with the general northeasterly strike of rocks and major faults and to the trend of the massif.

SLIGHTLY ALBITIZED AND AMPHIBOLIZED ROCKS

These rocks are unevenly distributed throughout the area. In the vicinity of the gabbroid intrusion, albitization is predominant, affecting a 150 to 200 m interval, chiefly of sandstones and siltstones. The albitized varieties acquire here a white to light-gray color, preserving at times traces of the original banding. They consist mostly (50 to 70%) of fine, 0.1 to 0.2 mm albite grains and of a small amount of quartz, chlorite, and calcite. The rutile content is somewhat higher (as much as 1%).

More common are albitized and amphibolized shales which form bands in the unaltered biotite shales. These bands are a few centimeters to a few decimeters wide and extend for several meters, less commonly for a few tens of meters. On the background of dark shales, these bands stand out in their white to greenish color.

The rocks consist of columnar (as much as 3 mm) actinolite grains ($c\gamma$ = 15°, 2V = 60°), and a small amount of quartz and albite, with impurities of K-feldspar, epidote, and calcite. Quartz and albite occur in small (0.2 to 0.3) isometric grains.

Orthoclase (γ = 1.525 and negative 2V of about 60°; $\angle\gamma\perp$ (001), 5°) occurs with actinolite, forming irregular, grownish in transmitted light, grains, as much as 0.2 mm, with sinuous outlines. In the replacement of biotite by actinolite, orthoclase usually is formed in the adjacent segments, most intensively at the

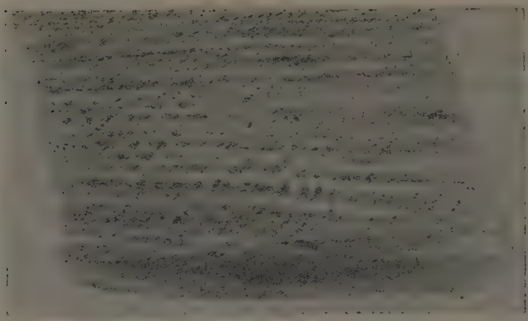

FIGURE 5. Amphibolitized and K-feldspathized rock and a secondary banding developed on cleavage. Four-fifths natural size.

boundary of actinolite segments with the biotite and quartz-feldspar, where it forms a reaction rim. These rims, 1 to 3 cm wide, are especially well developed along thin (as much as 5 cm) actinolite veins which cut the biotite shales.

The more intensively altered feldspar-amphibolite rocks are marked by a banded structure brought about by an alternation of light-colored quartz-feldspar and greenish, chiefly actinolite lentils. The banding usually coincides with the original bedding and its corresponding schistosity. However, locally it is at an angle to the latter, being developed upon the earlier stratification banding (Fig. 5). The chemical composition of this rock is given in Table 5.

In isolated instances, the feldspar-amphibole rocks were subject to later metasomatic changes which resulted in their nearly complete replacement by a white quartz-albite rock (Fig. 6) consisting of isometric to somewhat elongated grains of albite-orthoclase and quartz. The accompanying change in the composition is illustrated in Table V (XV and XVI).

Along with siltstones, dolomitic limestones, too, undergo some degree of albitization. The newly formed albite is represented there in euhedral prismatic grains, as long as 2 to 3 mm, isolated among carbonate grains of the granoblastic aggregate (Fig. 7). Albite in dolomitic limestones has developed exclusively at the expense of the influx of nearly all of its components.

In the Kodar Range area, metasomatic alterations affect Precambrian rocks which are stratigraphically correlative with the upper part of the Udokan series, the latter being more intensively metamorphised. Besides the strongly metamorphosed biotite sandstones, locally interbedded with limestones, crystalline schists and paragneisses are widely developed.

All these rocks make up the limb of a large anticlinal fold, with steep overturned, in places isoclinal folds of the second order. The rocks strike to the northwest, dipping northeast at 35 to 40°. Widely developed are zones of shattering, which strike to the northeast and northwest.

Dikes of light-colored granites and pegmatites, and quartz-feldspar veins are developed in the area.

FIGURE 6. Replacement of a feldspar-amphibole rock by a white quartz-feldspar rock. Natural size.

Table 5

The balance of matter in the change of amphibolized rock to the quartz-albite

Oxides	% by weight			% by weight reduced to 100			Weight of oxide in grams for 100 cm³ of rock			Transformation, change in oxide content, in g.
	XIV	XV	XVI	XIV-a	XV-a	XVI-a	XIV-b	XV-b	XVI-b	XV to XVI
SiO_2	63,76	86,62	86,75	63,69	86,59	86,84	170,05	232,94	227,52	—5,42
TiO_2	0,55	0,07	0,13	0,55	0,07	0,13	1,47	0,19	0,34	+0,15
Al_2O_3	13,80	6,96	6,95	13,78	6,95	6,96	36,79	18,71	18,24	—0,47
Fe_2O_3	0,80	0,16	0,31	0,80	0,16	0,31	2,14	0,43	0,81	+0,38
FeO	3,28	0,84	1,24	3,28	0,84	1,24	8,76	2,26	3,25	+0,99
MnO	0,04	0,01	—	0,04	0,01	—	0,11	0,03	—	—
CaO	5,40	0,57	0,30	5,39	0,57	0,30	14,39	1,53	0,79	—0,74
MgO	2,38	0,62	0,08	2,38	0,62	0,08	6,35	1,68	0,21	—1,47
Na_2O	7,75	3,55	3,38	7,74	3,55	3,39	20,67	9,55	8,88	—0,67
K_2O	1,18	0,44	0,36	1,18	0,44	0,36	3,15	1,18	0,94	—0,24
H_2O	0,18	0,12	0,08	0,18	0,12	0,08	0,48	0,32	0,21	—0,11
Loss in heating	0,99	0,08	0,31	0,99	0,08	0,31	2,64	0,21	0,81	+0,60
Total	100,11	100,04	99,89	100,00	100,00	100,00	267,0	269,03	262,0	—7,00
Volume wt.	2,67	2,69	2,62							

XIV -- albitized and amphibolized rock (V. M. Kovyazina, analyst); XV -- slightly amphibolized rock (V. A. Yusova, analyst); XVI -- lighter in color quartz-albite segments in XV (V. A. Yusova, analyst).

NOTE: Comma represents decimal point.

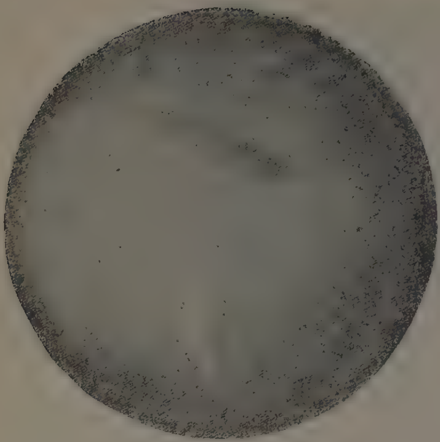

FIGURE 7. Development of albite porphyroblasts in dolomitic limestone. 15X; crossed Nicols.

Metasomatic changes occur chiefly in alternating metamorphic sandstones and limestones, mostly at contacts of different beds and along zones of shattering. As a result, amphibole-pyroxene-feldspar, albite, and actinolite-diopside rocks have been developed here.

The amphibole-pyroxene-feldspar rocks form lenses in skarn limestones, or else beds a few hundred meters thick and persisting to more than 1000 m. They differ from the enclosing limestones by their green color.

Most common are dense, fine-grained amphibole-feldspar rock, less commonly coarse-grained, mottled to banded. They are mixed-grained, usually granoblastic, serrate in structure, consisting of a quartz and feldspar aggregate with elongated amphibole grains. In mottled varieties, the light-colored segments, 1 to 2 cm, are made up chiefly of quartz and feldspar, with amphibole the best developed in this matrix.

Feldspars are usually represented by albite, oligoclase, and microcline, with albite developed after the oligoclase. Microcline either replaced the plagioclase or else was deposited in veins.

Amphibole is represented by actinolite and glaucophane. The actinolite forms coarse grains with a sievelike and skeleton texture, or else occurs in fine grains. It locally contains relicts of pyroxene and has been replaced by epidote. Glaucophane occurs in irregular, elongated grains displaying violet pleochroism. Biotite occurs in fine scales, along with a small amount of scapolite and carbonates. Accessory minerals are represented by tourmaline, apatite, sphene, and ore minerals.

Light-colored, fine-grained albite rocks with a granoblastic texture are less common. Besides the albite, their components include a small amount of quartz, oligoclase, microcline, actinolite, and calcite; with sphene, apatite, and ore minerals, for accessories.

Actinolite-diopside rocks occur in veinlike accumulations and in pockets of diopside and actinolite, but by a network of thin carbonate veins. Scapolite is common.

Contacts of mottled metasomatics with carbonate rocks commonly show a development of coarse-grained diopside-actinolite-scapolite rocks, as thick as 2 m.

Table 6 shows changes in rock composition, for a metasomatic process.

The formation of albite-actinolite rocks was accompanied by leaching of Al_2O_3, Fe_2O_3, and FeO; by a considerable leaching of K_2O; and by a considerable addition of Na_2O, CaO, and MgO. The development of albite rocks was accompanied by leaching of Fe_2O_3, FeO, MgO, and especially of K_2O, and by a considerable addition of Na_2O.

In isolated segments, metasomatic rocks include remnants of less altered rocks represented by metasandstones with a blastopsammitic texture, and by newly formed actinolite, scapolite, and albite.

We note, in conclusion, that isolated localities of this area also show an intensive microclinization, of a later origin than the widespread albitization: microcline is clearly developed on the albite. Of interest is the second albitization stage, which took place after the microclinization and was expressed in the formation of albite and albite-amphibolite veinlets and pockets in microclinic rocks.

CONCLUSIONS

1. Alkaline metasomatism of the Udokan series is expressed in a slight albitization, less commonly in biotitization of individual beds; or else in an intensive albitization, less commonly in biotitization of rocks in zones of shattering.

In the first instance, the metasomatic alteration had a frontal character, determined

Table 6

The balance of matter in the alteration of sandstones to albite-actinolite and albite rocks

Oxides	% by weight, reduced to 100			Weight of oxide in g, for 100 cm³ of rock			Change in the oxide content, in g, in the transformation	
	XVII-a	XVIII-a	XIX-a	XVII-b	XVIII-b	XIX-b	XVII to XVIII	XVII to XIX
SiO_2	62,46	61,75	65,74	166,14	164,25	170,92	— 1,89	+ 4,78
TiO_2	0,55	0,33	0,59	1,46	0,88	1,53	— 0,58	+ 0,07
Al_2O_3	15,34	11,61	15,80	40,80	30,89	41,08	— 9,91	+ 0,28
Fe_2O_3	3,13	0,30	0,30	8,33	0,80	0,78	— 7,53	— 7,55
FeO	3,11	2,05	1,08	8,27	5,45	2,81	— 2,82	— 5,46
MnO	trace	0,08	0,01	—	0,21	0,03	+ 0,21	+ 0,03
CaO	4,00	9,09	4,18	10,64	24,18	10,87	+13,54	+ 0,23
MgO	3,00	7,35	1,20	7,98	19,55	3,12	+11,57	— 4,86
Na_2O	2,07	5,18	6,91	5,51	13,78	17,97	+ 8,27	+12,46
K_2O	5,50	0,91	0,80	14,63	2,42	2,08	—12,21	—12,55
H_2O	trace	0,09	0,23	—	0,24	0,60	+ 0,24	+ 0,60
Loss in heating*	0,72	1,12	2,60	1,92	2,98	6,76	+ 1,06	+ 4,84
	0,12	0,14	0,56	0,32	0,37	1,45	+ 0.05	+ 1,13
Total	100,00	100,00	100,00	266	266	260	0,00	— 6,00
Volume wt.	2,66	2,66	2,60					

XVII -- hornfelsitic sandstone; XVIII -- albite-actinolite rock; XIX -- albite rock; all analyses by N. V. Lodochnikova.

*Russian original gives two lines of data for the one determination.

NOTE: Comma represents decimal point.

to some extent by lithologic features of altered rocks, with carbonate and carbonate-carrying rocks the main subject of the alteration. The albitization took place chiefly in siltstones and sandstones, locally calcareous, and to a considerably smaller extent in metashales. Metashales, dolomitic limestones, and magnetite-carrying intercalations were the main subjects of biotitization.

In the intensive metasomatism of shattered zones, the role of the original rock composition diminished, and the nature of metasomatism was determined mostly by the nature and strength of the altering solutions.

2. Alkaline metasomatism was accompanied by an extensive migration of the components. In places, a peculiar metasomatic differentiation of the latter took place, expressed in the enrichment of some rocks by their original components, present in them in excess. Thus, the albitization of siltstones and sandstones was accompanied by the addition of SiO_2, Al_2O_3, and Na_2O — the components originally present there in excess. The metasomatism of shales, with their high content in Al_2O_3, Fe_2O_3, and K_2O, was accompanied by their enrichment in these components. Thus metasomatic changes at times increased the original differences in the material composition of source rocks.

3. An analysis of the balance of individual components, in the process of alteration for different rocks, discloses that Fe_2O_3 and MgO were added in the biotitization of some rocks and leached out in their albitization. In view of the simultaneousness of these two processes, it may be supposed that they are interrelated in some way, so that the albitization of some rocks possibly involves the biotitization of the others.

This, however, does not mean that all processes of an alkaline metasomatism can be explained by such redistribution of the components.

The abrupt increase in the amount of alkalis in a sodium and potassium metasomatisms is undoubtedly related to their addition

by the reactive hydrothermal solutions.

4. It is possible that the coincidence in time of the sodium and potassium metasomatisms is a result of their relationship with a single altering front of hydrothermal solutions, whose change in character depended on the composition of the rocks being altered.

This may be especially well demonstrated in our instance, with the flyschlike character of the deposits expressed in an alternation of rocks extremely diversified in composition. This diversification determined the definite chemical anisotropy which persisted throughout the subsequent regional metamorphism. So it came about, that the transient hydrothermal solutions destroyed the pseudoequilibrium in the composition of these rocks, thereby determining the changes in the rocks themselves and the appearance of new hydrothermal components, considerably altering the nature of the solutions.

5. The time sequence for the several metasomatic processes has not been established. These processes affected the same zones, probably simultaneously, but involved rocks of different composition. No superposition of different alterations (namely, albitization and biotitization) in the same rock has been observed.

6. The manifestation of hydrothermal metasomatic processes within hornfels fields, in rocks previously undergoing contact metamorphism, suggests that most metasomatic alterations have not been the effect of a direct intrusive contact but rather a result of later hydrothermal processes, probably connected with that intrusion.

7. The processes of metasomatic alteration in Proterozoic and other rocks are very important as a controlling factor in ore formation.

These processes, active in zones which are tectonically weakened usually prior to mineralization, determine the appearance of metasomatically altered rocks, which are much different than the enclosing unaltered rocks, not only in their physical properties but also in composition. The subsequent pre- and intra-mineralization disturbances commonly take place within these altered rocks. The latter then become an arena for later, possibly ore-bearing, stages of hydrothermal mineralization.

To be sure, the specific chemical composition of these premineralization metasomatic formations played a definite part in the deposition of ores, whereas their textural and structural features affected the distribution of ore minerals in the rocks.

8. It is characteristic that ore veins which originated at later stages of the hydrothermal process, in places, have about the same composition of their component minerals as metasomatic rocks connected with an earlier stage of the same process. This is manifest in the above-named regions where quartz-feldspar and quartz-amphibole veins occur alongside the albitized and amphibolized rocks; this also occurs in other provinces and under other geologic conditions, where quartz-tourmaline veins are associated with earlier tourmalinized rocks, and chlorite veins are observed in previously chloritized rocks.

Thus it transpires that "near-vein" altered rocks are not always directly related to the formation of veins but rather are more complex formations, developed as a result of hydrothermal metasomatism, subsequent to the effect of contact metamorphism but prior to the appearance of ore veins and even predetermining it to a certain extent, in this or that locality.

In conclusion, we shall pause briefly for the problem of the genesis of alkaline metasomatism.

It should be noted that there is no unanimity of opinion on this subject. The authors who have studied the products of alkaline metasomatism voice quite different views on its source.

Yu. Ir. Polovinkina [4] emphasizes in her compilation, the spatial and genetic relationship of rocks, altered as a result of alkaline metasomatism, with Precambrian ferruginous quartzites developed in many places on the earth. She comes to the conclusion of a genetic relationship of the sodium metasomatism processes, as well as of the ferruginous quartzites themselves, with basic extrusive rocks.

On the basis of his observations in the area of the Kursk Magnetic Anomaly, D. S. Korzhinskiy [2] speaks of "the connection of alkaline metasomatism with the general metamorphism accompanied by the formation of normal alkaline granites." He explains the selective albitization of ferruginous quartzites by the sizable content of carbonates in the latter.

In explaining the important part of carbonate rocks in metasomatism, D. S. Korzhinskiy notes that the decomposition of carbonate-type salts with a weak base, leads to the increase of the overall alkalinity of the solution and to the increase in the activity coefficient for strong bases such as sodium and potassium.

In describing the phenomenon of sodium metasomatism in the area of Pokrovskoye

Village and in the northern part of Krivoy Rog, Yu. Ir. Polovinkina [5] relates the albitization processes with the action of hydrothermal solutions, subsequent to the pegmatites' formation. She notes that, in the Ukraine, only the pegmatites associated with the Korosten' complex are accompanied by sodium-rich solutions.

V.S. Dovarev [1] believes that alkaline metasomatism of the ferruginous Krivoy Rog quartzites is related to general metamorphic processes, and denies any part of igneous formations in this process. He believes that the metasomatism is determined by the solution and redeposition of substances from water liberated in the dehydration of rocks and minerals, in the course of diagenesis, folding, and metamorphism.

According to A.P. Nikol'skiy [3], alkaline metasomatism of the Krivoy Rog ferruginous quartzites was brought about by aplitic granites whose formation was separated by a considerable time interval from that of older basic intrusions. The alkaline metasomatism affects not only the ferruginous quartzites, but the basic rocks as well. This is explained by the presence in both of sizable amounts of Fe and Mg, and by their deficiency in alumina. A factor favorably affecting the formation of albites is supposed to be the interaction of ultrabasic and ferruginous rocks with an acid magma, which commonly leads to the formation of alkali rocks.

In considering the above-mentioned manifestations of an alkaline, specifically the sodium, metasomatism, it should be noted that, in those instances, it has been developed under different geologic conditions.

In individual instances, a relation between the sodium metasomatism with basic intrusive rocks has been established. It has also been established that the metasomatic processes did not accompany all of the basic intrusions, but rather was materialized under certain specifically favorable conditions. In such instances, albitization was, in places, associated with the contact action of basic intrusions and resembled the formation of adinole rocks.

In a number of the above instances, the albitization processes were probably connected with granite intrusions, wherein the albite rocks were not formed at the intrusive contact but were developed rather at the expense of hornfelses.

In conclusion, the author takes this opportunity to express his deep gratitude to Professor Yu. Ir. Polovinkina for her reading of this paper.

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RELATION BETWEEN THE FORMATION OF MAGNESIAN SKARNS AND GRANITIZATION[1]

by

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The problem of the relationship between the magma and enclosing rocks is as interesting as it is far from solution. The study of hybrid igneous rocks which originate at the contact with enclosing rocks makes it possible to judge, in isolated instances, the effect of these enclosing rocks on the change in the composition of the intruding magma, and the study of the halo of metamorphosed rocks about the intrusion aids in the understanding of the character of altering solutions which emerge at the igneous stage.

The scope of problems concerning the interrelationship between the magma and the enclosing rocks broadened considerably after D. S. Korzhinskiy advanced the hypothesis of granitization as a magmatic replacement, where an immense significance is attached to penetrating magmatic solutions which facilitate the advance of the granitization front and determine the metamorphic and metasomatic alterations during the igneous stage.

The data used in this paper have been obtained from a study of the Zheleznyy Kryazh (Iron Range) contact-metasomatic iron ore deposit, eastern Trans-Baikal region, where the interaction between the granitoids and enclosing sedimentary rocks is widely developed.

The Zheleznyy Kryazh deposit is located at the contact of the Kutomar Variscan granitoids with lower Paleozoic sedimentary rocks. The metasediments form remnants of various sizes at the top of the Kutomar granitoids and are represented in their lower part by hornfelses, magnesian and calcareous skarns, and quartzites. In their upper part, they are represented by chloritic sandy metashales (Fig. 1).

These remnants, separated by granitoids, represent segments of the northwestern limb of a large anticlinal structure, whose southeastern limb is located far beyond the ore field and whose axial part is formed by the granitoids.

As disclosed by exploratory boreholes, the remnants of sedimentary rocks which form the anticlinal limb are traceable at a depth of more than 300 m, and the erosion in its axial part has uncovered granitoids at their highest elevations. This suggests that the anticlinal structure of lower Paleozoic sedimentary rocks has facilitated in some way the rise of the granite magma.

Granitoids form sharp intrusive contacts with the metasediments. Prior to the metamorphism, rocks of the lower member consisted of sandstones interbedded with sandy and calcareous shales, and of intercalations and lenses of dolomites. As a result of the metamorphic and metasomatic processes, shales of the sedimentary sequence changed to hornfelses of various composition; sandstones turned to quartzites; and most of the carbonate layers, to magnesian and calcareous skarns which subsequently underwent considerable mineralization.

As it appears from the study of the Zheleznyy Kryazh deposit, calcareous pyroxene-garnet skarns are postigneous, formed after the upper parts of magmatic bodies had solidified. This is clear from the replacement of granitoids by calcareous endoskarns. In isolated instances, magnesian skarns are replaced by the calcareous skarns; i. e., the latter are the younger formations [14].

It has also been determined that magnesian skarns are cut by dikes of granite-aplite and early diorite porphyries, which have been replaced to various extent by the calcareous skarn minerals. It follows that these dikes intruded after the formation of magnesian skarns and prior to that of calcareous skarns.

Magnesian skarns and diopside-plagioclase skarn rocks exhibit an injection effect of the granitoids in contact with them (Figs. 1, 2),

[1] O svyazi obrazovaniya magnezial'nykh skarnov s granitizatsiyey.

whereas the hybrid granitoids show a definite metasomatic magnesian skarn zonation: a granitoid-diopside or diopside-spinel zone -- forsterite or forsterite-spinel zone. It is characteristic that granitoids in contact with magnesian skarns show no evidence of metasomatic alteration.

nesian skarns were formed before the solidification of the magma; i. e., during the igneous stage.

Intrusive rocks of the Zheleznyy Kryazh are represented chiefly by porphyritic biotite-hornblende granites.

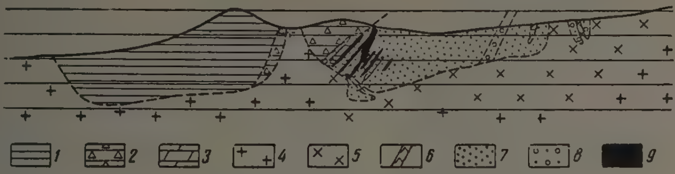

FIGURE 1. Cross section of the Zheleznyy Kryazh deposit. Scale, 1:10,000.

1 -- Siltstones and sandy metashales with subordinate sandstones; 2 -- quartzite and quartzite sandstones; 3 -- dolomites; 4 -- biotite-hornblende granites; 5 -- granites, granodiorites with subordinate syenite-diorites; the latter usually in the vicinity of magnesian skarns; 6 -- diorite porphyrites -- dikes; 7 -- hornfelses and relict layers of metashales and sandstones; 8 -- magnesian skarns; 9 -- ore and ore skarns.

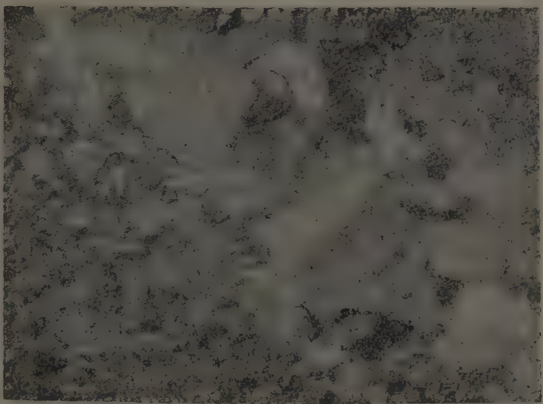

FIGURE 2. Diopside near-skarn rock cut by a monzonite vein.

Recrystallization and coarsening of diopside takes place along the vein walls. Thin section, X 16; Nicols crossed.

The definite metasomatic zonation of magnesian skarns about the granitoids suggest a genetic relationship between the two. The fact that both the near-skarn rocks formed on hornfelses, and the magnesian skarns, are cut by unaltered magmatic veins (Fig. 2), in the absence of endocontact metasomatic alteration in the granitoids, shows that the mag-

The tendency toward a porphyritic texture in granitoids of the area, together with the presence of zonal plagioclase, suggests their hypabyssal -- in M. A. Usov's terminology [18] -- position at the moment of crystallization. The comparatively shallow formation depths for the Kutomar granitoids were suggested also by V. A. Melioranskiy [15],

V. N. Kozerenko [13], and other students of the Trans-Baikalia.

Going toward the contacts with remnants of metasedimentary rocks, the porphyritic biotite-hornblende granites gradually change to granodiorites, syenites, syenite-diorites, and syenites. They all appear to be hybrid rocks which have originated from the interaction between a granite magma and the enclosing sedimentary rocks. This is especially obvious in the gradual transitions from granites to syenite-diorites. The best developed haloes of hybrid rocks have been observed at the intrusive contacts with magnesian postdolomitic (apodolomitic) skarns. The term, hybrid rock, in this instance, points to a dual origin of the rock substance, without saying anything of the method of its formation.

Macroscopically, the change in the normal granite composition, going toward the contact with the enclosing metamorphosed rocks, is expressed in a lower quartz content and in a higher content of feldspars and dark minerals. Gray plagioclase becomes predominant in porphyritic separates, and the rock itself becomes gray in contrast to the pink of the granites.

In isolated instances, coarse (as large as 4 x 4.5 cm) crystals of pink K-feldspar, standing out on the gray background of hybrid granitoids, have been observed near the latter's contact with metamorphic rocks which enclose the magnesian skarns. Under the microscope, monzonites and syenite-diorites show some diopside which is everywhere absent in granites. A study of syenite-diorites and monzonites reveals that diopside participates in the rock microstructure, as does K-feldspar which forms a granophyric texture with quartz. Thus, both these minerals are igneous rather than metasomatic in this instance.

Samples of igneous rocks have been analyzed in order to trace the changes in the chemical composition of nascent hybrid grani-

Table 1

Chemical analyses of igneous rocks, magnesian skarns, and dolomite from the Zheleznyy Kryazh deposit (in %)

Chemical components	Granite*	Granite	Granite	Grano-diorite*	Syenite	Mon-zonite	Syenite-diorite	Magnesian, partially serpentinized forsterite-spinel skarns		Dolo-mite
	anal. 1, spec. 927	anal. 2, spec. 1939	anal. 3, spec. 549	anal. 4 spec. 18/41	anal. 5 spec. 1153	anal. 6, spec. 497	anal. 7, spec. 24–102	anal. 8, spec. 1029	anal. 9, spec. 483	anal. 10, spec. 914
SiO_2	72,10	71,15	72,13	66,88	51,68	60,10	66.95	35,74	30,56	1,28
TiO_2	0,10	0,16	0,29	0,40	0,41	0,71	0,42	0,27	0,24	—
Al_2O_3	14,03	14,56	14,11	16,24	18,39	17,63	15,14	8,30	8,81	0,25
Fe_2O_3	1,90	1,59	1,12	2,95	0,96	2,65	0,46	5,91	5,78	2,37
FeO	0,93	1,43	0,85	0,86	4,72	2,62	1,43	3,93	2,84	0,20
MnO	0,04	0,05	0,43	0,09	0,07	0,52	0,34	—	—	—
MgO	0,41	0,08	0,51	0,90	9,66**	2,49	1,44	33,33	36,03	20,83
CaO	0,58	1,81	1,31	2,54	1,74	5,34	3,13	3,59	0,62	31,70
Na_2O	3,72	3,64	3,13	3,70	4,02	4,83	2,62	0,34	0,47	0,10
K_2O	4,62	4,77	6,18	3,85	3,24	2,69	7,79	0,26	0,40	—
H_2O^+	0,04	0,03	0,07	0,20	0.38	0,30	0,12	0,58	1,05	0,18
H_2O^-	0,29	0,26	0,57	0,37	3,74	0,42	0,33	7,29	11,68	3,12
CO_2	—	—	—	—	—	—	—	0,25	1,0	39,54
ZnO	—	—	—	—	—	—	—	0,28	0,30	—
S	—	—	—	—	—	—	—	—	0,22	—
Total	99,62	100,24	99,40	100,32	99,37	99,87	99,90	100,59	100,23	99,65

* Analyses 1 and 4 are borrowed from an unpublished paper of I. A. Yefimov.
** High content because of late chloritization.

Note: Comma represents decimal point.

toids, caused by interaction of the granite magma with sedimentary rocks, across the zonal transition from granites to magnesian skarns. The results are given in Table 1, along with the analyses of nonreplaced dolomites and of magnesian skarns which have developed at their expense. Data of Table 1 are partially illustrated by variation diagrams (Fig. 3).

The following hybrid rock series, from granites to magnesian skarns, occur within the ore field (the hybrids are underlined):

1. Granites (analysis 1) -- granites (analysis 2) or granodiorites (analysis 4) -- syenites (analysis 5) -- magnesian skarns (Zheleznaya Mountain area).

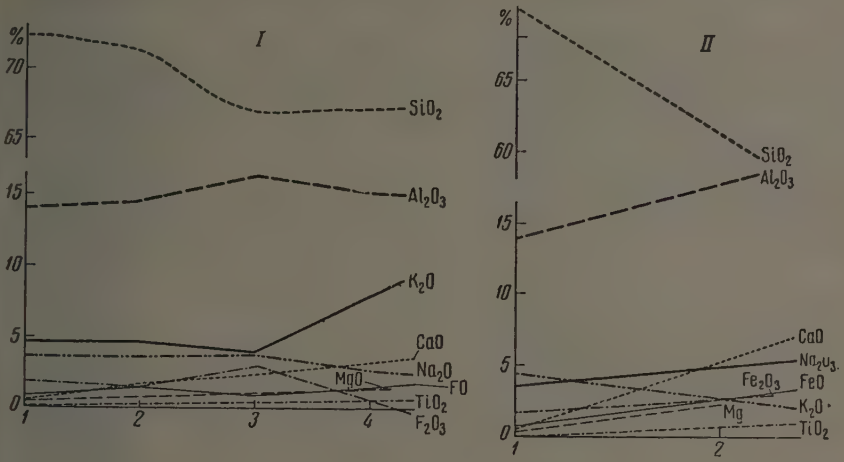

FIGURE 3. Variation diagrams for the chemical composition of granitoids.

I. 1 -- granite (analysis 1), away from contact with sedimentary rocks; 2 -- granite (analysis 2); 3 -- granodiorite (analysis 4) near contact with sedimentary rocks; 4 -- syenite-diorite, directly at contact with a magnesian skarn;

II. 1 -- granite (analysis 1); 2 -- monzonite from contact with a magnesian skarn.

Table 2

Numerical characteristics of igneous rocks from the Zheleznyy Kryazy deposit, recomputed by the A. N. Zavaritskiy method

Anal. Nos. Table 1	a	c	b	S	f'	m'	a'	c'
1	14,3	0,8	4,9	80	51	13	36	—
2	14,5	2,1	4,7	78,7	66,7	29	4,3	—
3	15,3	1,5	3,2	80,0	71	25,8	—	3,2
4	13,8	3,1	6,8	76,2	48	24	28	—
6	14,7	4,6	10,9	69,8	43,7	38,6	—	17,7
7	16,8	1,6	7	74,6	34,3	34,3	—	31,4

Note: Comma represents decimal point.

2. Granites (analysis 1) -- granites or granodiorites-syenite-diorites (analysis 7) -- magnesian skarns (the Main [Glavnaya] deposit area, see diagram I, Fig. 3).

3. Granites (analysis 1) -- monzonites (analysis 6) -- magnesian skarns (Pryamaya Rapids watershed, see diagram II, Fig. 3).

4. Granites (analysis 1) -- granites of higher alkalinity (analysis 3) -- magnesian skarns (Shirokiy Log area).

A correlation of the chemical analyses (Table 1, Fig. 3) and the numerical characteristics of granitoids (Table 2) shows the following changes in their composition, going toward the skarn contact: a decrease in the quartz content (quantity s), a decrease in the basicity of plagioclase (c), a rise in the colored components content (b), an accumulation of magnesium in the colored fraction, and an increase in the alkali content near the contacts.

To determine the nature of the process responsible for the formation of the Zheleznyy Kryazh deposit, it is important to know whether the origin of hybrid granitoids was due to an assimilation process in the enclosing rocks or to their granitization. Before attacking this problem, it is necessary to clarify the meaning of the terms used.

Although the term, assimilation, is on the whole generally understood in a single sense, this is not true for granitization. Although, in the final reckoning, this term designates the changes of sedimentary and extrusive rocks to granitoid-type rocks, the process itself is interpreted differently by different geologists.

Specifically, H. Reid [16] regards granitization as "a process wherein solid rocks are turned into granites, without passing through a magmatic stage;" i. e., the transformation of various solid rocks to granitoids is effected by metasomatic processes. This concept of granitization has found support in a number of papers of G. D. Afanas'yev [1, 2], although he assigns a subordinate part to metasomatic processes, as compared with the igneous, especially under hypabyssal conditions.

The possibility of the granite formation by metasomatism is mentioned by N. G. Sudovikov [17], N. M. Uspenskiy [19], and other Soviet and foreign geologists.

Other geologists, for example, D. S. Korzhinskiy [8] regard granitization as an infiltrational magmatic replacement of the enclosing rocks, with the formation of granitoids. In other words, according to them, granitization must pass through an intermediate stage of solution or melting of the enclosing rocks, with the overall composition of the melt approaching the eutectic for granitoids.

Inasmuch as distinct cutting intrusive contacts between granitoids and metasedimentary rocks have been observed everywhere in the Zheleznyy Kryazh deposits, in the absence of any evidence as to a metasomatic origin of the granitoids, we apply the term, granitization, in this paper to mean magmatic replacement as understood by D. S. Korzhinskiy.

One of the features of the granitization phenomenon, in the Zheleznyy Kryazh deposit, is the formation of infiltration-type magnesian skarns, during the igneous stage.

The formation of magnesian skarns during the igneous stage; i. e., before the magma had time to solidify, points to an intensive circulation of solutions related to the granitoid intrusion. D. S. Korzhinskiy [8] has named them, diamagmatic (penetrating magmatic). Flows of such solutions are of an infiltrational type.

The infiltrational character of diamagmatic solutions active in the Zheleznyy Kryazh deposit, is proven by the formation of such minerals as spinel; i. e., minerals originating under the conditions of a high chemical potential for Al_2O_3. The formation of spinel on dolomites is inconceivable in a merely diffusional displacement of the components, when one component is displaced in the direction of its lower chemical potential. When spinel skarns are formed on dolomites, the chemical potential of Al_2O_3 turns out to be higher in exocontact rocks than in the granitoids themselves. Consequently, the replacement was of an infiltrational rather than of diffusion character. The displacement of magnesium oxide in the magnesian skarn zones, as well as the leaching of calcium oxide from these zones, also suggests an infiltrational character of diamagmatic solutions. The Zheleznyy Kryazh hybrid granitoids, formed near the magnesian skarn contact, are characterized by their high alkalinity, despite their varied composition (see Table 1). Clinopyroxene-orthoclase paragenesis has been observed in syenite-diorites and monzonites. According to D. S. Korzhinskiy [12], this suggests a rise in alkalinity of the magma during the formation process for these rocks.

The formation of high alkalinity rocks would be difficult to explain by the assimilation of dolomites by granite magma. Assimilation is a diffusion process of solution of the enclosing rocks in the magma. Inasmuch as potassium and sodium are very mobile elements as compared with iron, aluminum,

etc., their chemical potentials would have the time to equalize along the assimilation front; i.e., along the diffusion displacement of the main component of the melt [10, 11]. In the Zheleznyy Kryazh deposit, however, high-alkali rocks occur at contacts with postdolomitic (apodolomitic) magnesian skarns.

All this leads to the conclusion that near-contact high-alkali rocks are formed in the process of granitization of dolomites, rather than of their assimilation.

In the presence of postdolomitic (apodolomitic) skarns in the granitoid field, and taking into account all the evidence of magmatic granitization, it is reasonable to assume that hornfelses and metashales which earlier enclosed dolomitic lenses, as well as partially the dolomites themselves, have been replaced by the granitoids. In other words, they were fused, and the granitoids, now in contact with skarns, have crystallized out of the hybrid magma.

It appears that the initial intrusive stages witness the interaction of granite magma with the unmetamorphosed enclosing rocks -- dolomites and shales -- which then are changed to various metamorphosed rocks, as an effect of circulation solutions of the granite magma. The further interaction of the granite magma, in the progress of the granitization front, took place with metamorphism products of the enclosing rocks. The solution and replacement of various enclosing rocks by the granite magma produced diverse effects on the composition of the resulting granitoids.

The replacement of sandy shales and sandstones, and of the products of their metamorphism, does not appear to cause extensive changes in the chemical composition of granitoids. This is because granites, in contact with those rocks, usually have a standard composition. On the other hand, the interaction of granite magma with carbonate-argillaceous rocks, dolomites and the products of their metamorphism, unavoidably leads to a change in the composition of the resulting granitoids.

An idea can be gained of the replacement of enclosing rocks by granitoids from a study of the replacement of xenoliths common in the Zheleznyy Kryazh hybrid granitoids.

It appears that, as an effect of magmatic melt, xenoliths are broken into smaller fragments, commonly threaded through by granitoid veins. Ini isolated instances, almost all of the xenolith is replaced by granitoids, with only a small accumulation of dark-colored minerals for a vestige of the unreplaced rock (Fig. 4).

Broken hornfelses, in the process of replacement by monzonites, occur in the Pryamaya Rapids -- Kulinda watershed. Under the microscope, they are exceedingly fine fragments of biotite-plagioclase hornfels, remnants of coarser xenoliths and preserving their hornfelsic structure (Fig. 5).

In individual crystalline andesine grains, in monzonites (No. 35), there are inclusions of plagioclase grains from hornfelses with a more basic composition (No. 50) (see Fig. 6).

In this area, magnesian skarns occur at

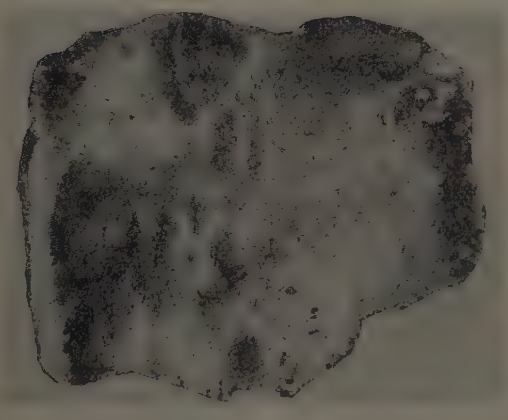

FIGURE 4. Replacement of xenolith (dark) by granitoid (light).

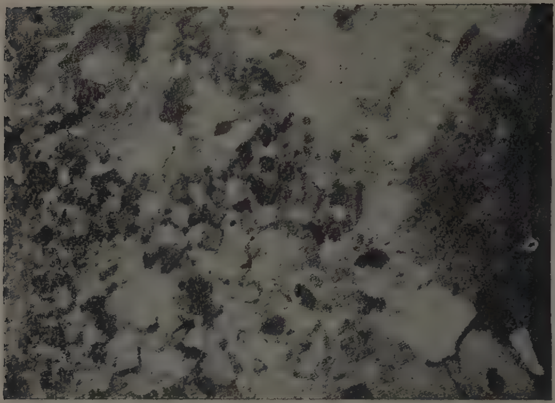

FIGURE 5. Segments of intact biotite-plagioclase hornfelses in granitoid, specimen No. 497. Thin section; X 37, crossed Nicols.

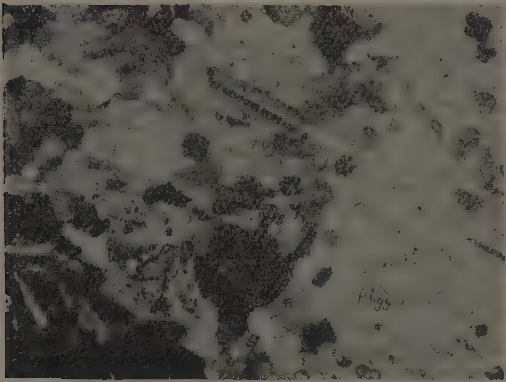

FIGURE 6. Fine plagioclase grains from hornfels are included in andesine from granitoid specimen No. 497. Thin section; X 56, crossed Nicols.

the contact with monzonite. It appears that dolomites and partially the magnesian skarns, as well as hornfelses, have been replaced by granitoids, in the process of granitization. Consequently, the basicity and alkalinity of the granite magma increased until hybrid magma of a monzonite composition was formed.

The Zheleznyy Kryazh magnesian skarns rest on all sides, as for instance southeast of the main deposit, the Zheleznaya Mountain, Shirokiy Log, etc. Preserved with the magnesian skarns, there are here thin hornfelses formed on shales which earlier carried dolomite lenses. No dolomite relicts have been found in those localities.

Let us look now into the stability of dolomites at the contact with granite magma.

An idea of the temperature of granite magma can be gained from the well-known experimental data on the fusing of igneous rocks. For instance, J. Douglas [22] obtained a free flow of diorite at 1,125° C; of granite, at 1,255° C; the beginning of fusing for syenite, at 1,165° C (at a pressure of one atmosphere). The experimental data of R. Goranson [23, 24] and N. Bowen and O. Tuttle

[21] yield lower temperatures for the fusing of granite, which is related to the water content in the melt. Thus, according to R. Goranson, water-free granite (at 1 atm.) has a fusing point of about 1,050° C. With the water content increased to 6.5%, and the water vapor pressure of 1,000 atm., the temperature of the full fusing of granite is lowered to 700 ± 50° C. At a lower content of water, and apparently of other volatiles, the fusing temperature of a melt rises. According to R. Goranson, the fusing temperature for granite with the water content at 2%, is 1,000° C.

What happens then to a dolomite in contact with granite magma at a temperature below 1,000° C? For a reply to this question, let us turn to some experiments with carbonate rocks.

According to experimental data of A. I. Tsvetkov [20], at a temperature of 740° C and pressure of 1 atm., dolomite dissociates into calcite and magnesite, according to the equation

$$CaMg(CO_3)_2 \rightarrow CaCO_3 + MgCO_3.$$

This is followed by a vigorous decomposition of magnesite into periclase and carbon dioxide, because magnesite, according to A. I. Tsvetkov [20], dissociates, at 580-630° C and 1 atm., according to the equation $MgCO_3 \rightarrow MgO + CO_2$.

The dissociation of calcite, at CO_2 pressure of 1 atm., takes place only at 898° C.

$$CaCO_3 \rightarrow CaO + CO_2.$$

These temperature data for the carbonate dissociation correspond to the carbon dioxide pressure of one atmosphere. At a depth of several kilometers below the surface, where the formation of the intrusive and the ore deposit took place, this pressure undoubtedly was different.

D. S. Korzhinskiy [7] points out that the average CO_2 pressure at depth equals the hydrostatic rock pressure in the absence of faults leading to the surface. Thus, at a depth of 4 km, the CO_2 pressure will be about 1,000 atm. The computed data on the metamorphism of carbonate rocks at different temperatures and pressures are cited in a number of papers of D. S. Korzhinskiy [6, 7].

There are experimental data on the same subject by R. Harker and O. Tuttle [25]. Given below is the pressure-temperature curve for dolomite dissociation, obtained by them for different temperatures and CO_2 pressures. This curve gives an idea of the phenomena taking place in dolomites at contact with magma at different CO_2 pressures corresponding to different depths, and of various values of temperature corresponding to the degree of heating of the dolomites by granite magma. For instance, according to the curve (Fig. 7), dolomites do not dissociate at the depth of 3 km, which corresponds to CO_2 pressure of 750 atm. and temperature of 700° C; they do so, however, at 800° C, breaking into calcite, periclase, and carbon dioxide. With the increase of the CO_2 pressure, the 800° C temperature is inadequate for the dolomite dissociation; the latter again proceeds when the temperature rises above 800° C.

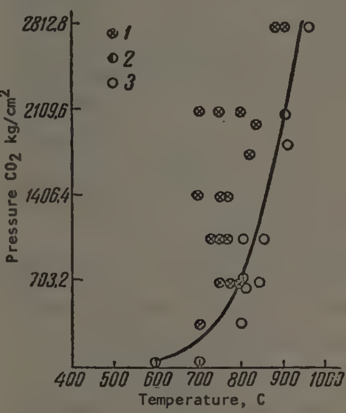

FIGURE 7. Pressure-temperature curve for dolomite dissociation.

1 -- dolomite; 2 -- dolomite + calcite + periclase; 3 -- calcite + periclase.

It follows from the data of R. Harker and O. Tuttle that at 1,000° C the dissociation of dolomites will proceed at the CO_2 pressure of 2,800 atm., which corresponds to a depth of about 12 km (see Fig. 7).

Keeping in mind the comparatively shallow origin of the Kutomar granitoids, we can assume this depth as no more than 10 km, in which case the CO_2 pressure did not exceed 2,500 atm.

It follows that dolomite lenses of the Zheleznyy Kryazh, in immediate contact with granite magma, could not have remained unmetamorphosed. At depths shallower than 10-12 km; i.e., under the conditions of a lower CO_2 pressure, dolomite dissociates at temperatures lower than 1,000° C. Limestones, on the other hand, do not dissociate at these temperatures and at high CO_2 pressures. However, a high content of water and

other volatiles in a granite melt is conceivable, in which case its temperature might be lower than 1,000° C (experiments of Goranson and others). In those instances, because of the action of water solutions on the enclosing rocks, the temperature of the solution, the fusing, and the metamorphism of the latter will be correspondingly lower.

The formation of hybrid granitoids at the expense of carbonate rocks demonstrates that the latter are dissolved, partially enriching the magma with CaO and MgO bases, although the temperature of the melt is likely to be below that of Ca fusing (1,339° C; 26).

In Zheleznyy Kryazh, near the contact with postdolomitic (apodolomitic) magnesian skarns, the nascent hybrid granitoids contain considerably more CaO and MgO than the granites (see Table 1, Fig. 3). The conclusion is that the action of high-temperature solutions on the enclosing rocks lowers the temperature of their fusing and metamorphism, as is the case of granites under experimental conditions.

The newly arriving batches of diamagmatic solutions carry the excess bases from the hybrid magma. Because of the differential intensity of these solutions, hybrid magmatic rocks of different composition are formed at different contacts.

The liberation of MgO as a result of the lower dissociation temperature of magnesite as compared with calcite, at high CO_2 pressures, appears to give priority to MgO over CaO, in their reaction with the components of diamagmatic solutions (SiO_2, Al_2O_3), which is what leads to the formation of the magnesian skarn minerals.

The formation of magnesian skarns with calcite, and of calcifers, is possible with a sluggish circulation of diamagmatic solutions through the dolomites. Such calcifer lenses, changing to dolomites, occur here away from the granitoid contacts. They usually are isolated from the latter by hornfels layers which interfere in some way with the flow of diamagmatic solutions. An intensive circulation of the latter flushes practically all of the CaO out of the magnesian skarn zone. This is probably what took place in the Kheleznyy Kryazh where dolomites form small bodies, during intensive granitization (see analyses in Table 1).

The initial granitization stage for dolomite lenses undoubtedly was marked by the appearance of zones where magnesian skarn minerals and calcite were formed after the dissociation of dolomite and magnesite. Then, as the granitization front advanced, and as new batches of diamagmatic solutions constantly flowed through the skarn zones, a partial solution of earlier-formed magnesian minerals took place, along with the displacement of MgO into zones where it could again react with the components of diamagmatic solutions. In the process, MgO-replaced calcite, which then was partially or fully dissolved and carried out of the magnesian skarn zones. This is confirmed by the microscopic study and chemical analysis of the Zheleznyy Kryazh magnesian skarns (analyses 8 and 9, Table 1).

As studied on the magnesian skarn zones of the Zheleznyy Kryazh, the metasomatic column can be presented as follows: 1) magma; 2) pyroxene or pyroxene-spinel zone; 3) forsterite or forsterite-spinel zone; 4) forsterite calcifer zone; and 5) dolomite.

This zonation is very similar to that described by V. A. Zharikov [3] from the western Kara-Mazar polymetal skarn deposits, where he first studied in detail the formation conditions for magnesian skarns.

As the granitization front advanced, the magma replaced the pyroxene zone, which in turn had been developed from the forsterite, and the latter from the calciferous. Because of the small thickness of dolomite lenses and of the intensity of the granitization, the Zheleznyy Kryazh skarn bodies exhibit two zones: a pyroxene or pyroxene-spinel, and a basic forsterite or forsterite-spinel.

Because the quartz-pyroxene-feldspar and biotite hornfelses, which enclose the magnesian skarns, are more readily fused than forsterite-spinel rocks, fragments of magnesian skarns with hornfels intercalations are preserved in many places in a granitoid field.

In citing this extensive evidence for a widespread granitization in the Zheleznyy Kryazh deposit, we do not mean to reduce to this process all of the interaction between the granite magma and the enclosing rocks.

The flows of infiltrating diamagmatic solutions, which determined the intensive granitization of enclosing rocks, undoubtedly did not present a solid front along the entire contact of magma and the enclosing rocks.

A localization of these flows appears to have been determined to a considerable degree by various geologic and structural factors. Among them were folding which facilitated the rise of magma, the zones of major faults, the presence of comparatively readily fusible and soluble rocks, etc.

It is likely that the zones or direction of an intensive granitization of the enclosing

rocks contained other zones where the interaction of magma and enclosing rocks was most likely reduced to a diffusion exchange of the components. As a result, the assimilation phenomena and the formation of highly basic hybrid rocks took place at the contact of the granite magma with carbonate-argillaceous enclosing rocks. In the Zheleznyy Kryazh deposit, such rocks include granodiorites and granites, which approach them in their composition but lack their higher alkalinity.

SUMMARY

1. The Zheleznyy Kryazh magnetite deposit exhibits a development of magnesian and later calcareous skarns, which originated from dolomites and other sedimentary rocks making up large remnants in the Kutomar granitoid massif.

2. Magnesian skarns of the area, unlike the calcareous skarns, were formed during an igneous stage, in the action of diamagmatic (after D. S. Korzhinskiy) solution of a granite magma, on dolomites.

3. The infiltration-type diamagmatic deposits determined the processes of the dolomite granitization. As a result, along with the formation of magnesian skarns in exocontacts, granitoids of a hybrid composition and of a higher basicity and alkalinity, originated in the endocontact zone of granites.

4. The higher alkalinity of endocontact granitoids is related, in this instance, to the higher activity of alkalis in diamagmatic solutions which is due in turn to their interaction with dolomites; i. e., with rocks marked by a high concentration of bases [D. S. Korzhinskiy, 9, 10].

This higher alkalinity makes it possible to distinguish a magmatic infiltration replacement -- granitization -- from the assimilation phenomena wherein, as a result of diffusion processes, an equalization of the chemical potentials for the alkalis took place, and the formation of high alkalinity rocks became impossible.

5. Besides the dolomites, the magmatic replacement (granitization) also affected large bodies of hornfelses which had originally carried the dolomites, and had been represented by sandy shales, the remnants of the enclosing rocks in the Kutomar granite massif.

6. Along with the granitization of dolomites and other enclosing rocks, the Zheleznyy Kryazh deposit witnessed the assimilation of the latter by granite magma, with the formation of highly basic granitoids. These phenomena took place along those segments of intrusive contacts where, because of their lithology and structure or for some other reason, the infiltration diamagmatic solutions were unable to circulate.

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SOME FEATURES OF THE ORIGIN OF THE URUP PYRITE DEPOSIT (NORTHERN CAUCASUS)[1]

by

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In the last decade, commercially important copper ore deposits have been discovered in the northern Caucasus, the most interesting of them being the Urup group of pyrite deposits. The following material mades it possible to determine their age.

THE PREORE AGE OF DIKES

The Urup pyrite deposits are made up chiefly of middle and upper Paleozoic rocks, also some Mesozoic rocks, all of which differ from each other in lithology, occurrence, and the degree of metamorphism.

Devonian rocks have undergone here a regional metamorphism, acquiring as a result a greenstone aspect and definite schistosity. Upper Paleozoic and Mesozoic deposits have not become schistose and are comparatively fresh looking.

The Urup intrusives are represented by sills and dikes of quartz diorites, lamprophyres, and diorites, fresh looking and massive in texture, within the metamorphosed sequence (Fig. 1).

Intrusive bodies have not been observed in upper Paleozoic and Mesozoic rocks of the Urup deposit area. Pebbles of nonschistose diorites occur in a middle Carboniferous basal conglomerate (unpublished data of Pilyuchenko and Kraus). This means that the dike intrusion is pre-middle Carboniferous but subsequent to the regional metamorphism of the enclosing rocks.

Unlike the dikes of quartz diorites and diorites of the area, dikes within the ore bodies have undergone intensive ore-contact alterations, expressed in quartzitization, sericitization, chloritization, and carbonatization.

The amount of pyrite increases in near-wall part of dikes, and at the contact with the ore where the altered dike rock is represented essentially by chlorite. In addition, chlorite is developed in a halo about the preore faults which cut the dike. The chlorite segments in the immediate vicinity of the preore faults or of the wedging-out parts of the dike usually carry pocketlike accumulations of metasomatic quartz (Fig. 2), locally with chalcopyrite veins.

The more intensive alteration of dike rock toward the ore contact, also near faults which cut the dike and in the latter's wedge-out, suggests that the rock was altered by hydrothermal solutions which accompanied the mineralization. Chalcopyrite veins cut the altered lamprophyre, their number increasing toward the periphery where they are oriented most commonly parallel to the dike-rock contact and are in accordance with the fracturing trend, in that segment. In the middle parts, where the fracturing is less intensive and not in conformity with the trend of the dike, the chalcopyrite veins are less numerous and irregularly oriented, in most instances. The similarity in the orientation of fractures and chalcopyrite veins in different parts of a dike also suggests a migration of mineralizing solutions to rocks with an already definite orientation of the fractures; i.e., pre-ore age of the dikes. Preore thrust faults which cut the dike are commonly filled with chalcopyrite, with a broad halo of an intensively chloritized rock about them. In a massive ore, the preore fault planes are difficult to trace.

Thus, the main reasons for a preore age of the dikes in the Urup pyrite ores are, briefly, as follows:

1. Only those dikes which cut the pyrite ore are intensively altered.

2. The intensity of alteration for the dike rock increases at the ore contact.

3. The orientation of fractures in different parts of a dike is identical with that of the

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[1]Nekotoryye cherty genezisa krupskikh kolchedannykh mestorozhdeniy (severnyy Kavkaz)

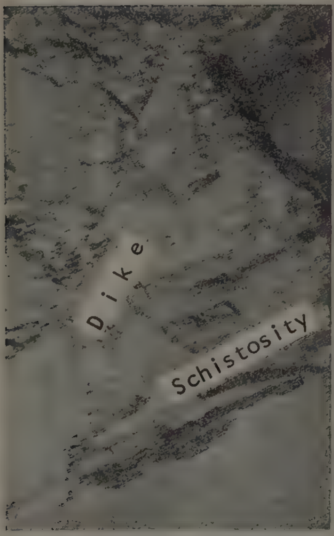

FIGURE 1. Dike of an altered vein rock, cutting schistose rocks. Skalistoye (Rocky) pyrite deposit. 10X.

chalcopyrite veins within the dike.

4. Thrust vaults which cut dikes are pre-ore, because their planes are filled with chalcopyrite, with the halo of an intensive hydrothermal alteration about them; the fault planes are not traceable in a massive ore.

A thin dike of altered lamprophyre cuts the dispersed copper-pyrite ores of the Skalistoye deposit at an angle to the schistosity of the rock (see Fig. 1). The alteration in the dike is expressed by intensive quartzitization. Its rock contains, in places, coarse incrustations of pyrite and chalcopyrite.

Within the ore body, chalcopyrite commonly fringes the dike with veins, 4 to 8 mm thick. They penetrate the dike along calcite veins which cut both the dike and the schistose rocks. Thus, the dike in dispersed ores of the Skalistoye deposit has also undergone ore-contact alteration and is therefore preore.

The following conclusions can be derived from the age of the dikes and from their preore origin: all of the known dikes in the Urup pyrite-deposits area are nonschistose; they are represented by fresh-looking rocks; they cut the enclosing schistose rocks; and they were formed after the regional metamorphism of the latter. The near-ore alterations in dikes within the ore bodies suggest that the mineralization is younger than the dikes.

AGE OF THE ORE BODIES, RELATIVE TO THAT OF THE ROCK SCHISTOSITY

The hanging wall of the Urup ore body is represented by quartzitic rocks formed in a zone of shattering from red siliceous schists,

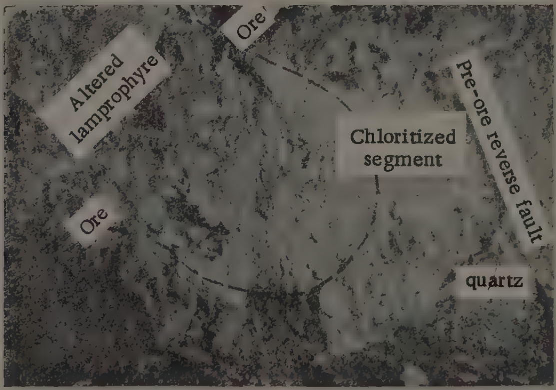

FIGURE 2. Preore reverse fault displacing the dike.

Near-ore replacement of lamprophyre by chlorite, more intensive at the thrust; accumulation of metasomatic quartz between dike and fault. Urup deposit.

FIGURE 3. Preore breccia of schistose siliceous shales.

Breccia cemented with massive pyrite ore. Slaty partings in the breccia fragment are filled with quartz and pyrite. 2X. Urup deposit.

FIGURE 4. Ore veins displacing quartz veins which unconformably cut a schistose rock. Vlasinchikha deposit.

with sulfide veins.

The presence of fragments of schistose rocks in the body of secondary quartzites indicates that the shattering affected rocks already regionally metamorphosed; the massive aspect of the secondary quartzites which cement these fragments also suggests that the breccia was formed after the advent of regional metamorphism.

The secondary quartzite zone is enriched by veins of pyrite and chalcopyrite. Their number is usually greater at the floor of the zone where they form a nearly massive body of pyrite ore. Thus it can be stated that the ore was formed after the regional metamorphism. This conclusion is also corroborated by the presence of a breccia of schistose siliceous shales, cemented with a massive pyrite ore (Fig. 3).

The hanging wall of the Vlasinchikha ore body locally exhibits a tectonic breccia of the overlying schistose tuffs of quartz albitophyres with similar features.

The enclosing rocks near the Urup ore bodies carry veins of pyrite and chalcopyrite which cut unconformably the schistose rocks. Their boundaries are sharp and clean cut; the ore texture is mostly massive.

The conformable pyrite veins, approximately parallel to the schistosity, offset the quartz veins which also cut the schistose rocks (see Fig. 4). Consequently, the formation of pyrite veins took place after that of the quartz veins and after the emergence of schistosity in the enclosing rocks.

These data lend substance to the assumption that the ore veins and the ore bodies with them, are of the same origin, having been formed after the regional metamorphism of the enclosing rocks.

The dependence of near-ore alterations on the textural features of enclosing rocks has been expressed in several ways. Siliceous metashales, in the vicinity of ores were chloritized: a black-green fringe of chlorite with a small amount of pyrite was formed at the immediate contact with the ore body. The upper part of this fringe exhibits gray-green bands of chloritized siliceous metashales alternating with thin bands of chlorite and quartz along the slaty partings, and gray-green chloritic siliceous metashales display red streaks of unreplaced siliceous metashales. Green-gray bands of altered siliceous metashale, near the contact, thin out gradually and are finally replaced by thin bands of quartz with chlorite. These data bear testimony to the fact that near-ore alterations affected rocks already schistose.

THE SO-CALLED "METAMORPHIC VEINS"

The Urup pyrite ore body is definitely asymmetric, with iron pyrites in the floors, and the copper pyrites and copper-zinc ores in hanging walls, where copper pyrite ores occur chiefly above the copper-zinc. Ore bodies in thin segments are commonly made up of a single ore; in those instances, they are dissimilar also in their content of useful components.

The Urup ore body is locally fractured, with a thin chalcopyrite fringe lining the fractures. The central parts of the fractures are nonmineralized, or else carry a small amount of fine calcite crystals. Such fractures occur along faults which cut the schistose rocks.

The ore body also contains thin chalcopyrite veins which cut both the quartzitic zone of the ore body and the overlying siliceous metashales. Chalcopyrite of these veins penetrates and fills fractures in the latter.

Chalcopyrite veins maintain their consistent thickness of 1-2 mm for a considerable distance, as much as tens of centimeters in a straight line. They were formed in cleavage fractures and have not been affected by the regional metamorphism.

ORE TEXTURES AS RELICTS OF THE REPLACED ROCKS TEXTURE

A proof of the replacement by ores of rocks in the hanging wall and floor is the presence of unreplaced mineral relicts typical of this or that rock.

Fine-grained microgranoblastic quartz, a product of crystallization of siliceous meta shales, occurs in pyrite ores of iron and copper of the Urup deposit which occur in the floor of the ore body.

Relicts of porphyrytic inclusions of quartz albitophyres occur in iron pyrite ores; i.e., nearer to the quartz albitophyres. They have not been found in the overlying varieties of ore. It follows that, in this segment of the ore body, iron pyrite, and partially the copper-pyrite ores, have been formed chiefly in the replacement of quartz albitophyres, whereas the copper pyrite varieties of massive ores have been formed in the replacement of siliceous metashales. The Urup ore body also displays an asymmetry in the position of typical textures in massive ores, which is in accordance with the above conclusion.

Uniform banding has been observed chiefly in copper-pyrite ores which occur, as a rule, in the hanging wall of the ore body, whereas the noncurved surface of slaty partings is typical of siliceous metashales in the hanging wall (Fig. 5). Curved banding is peculiar chiefly to copper-pyrite ores occurring usually in the floor of the ore body, and the curved surfaces of slaty partings, which fringe the schistosity lenses, are typical of quartz albitophyres in the floor (Figs. 6 and 7). This regularity proves that the ore has replaced the rocks in the hanging wall and the floor.

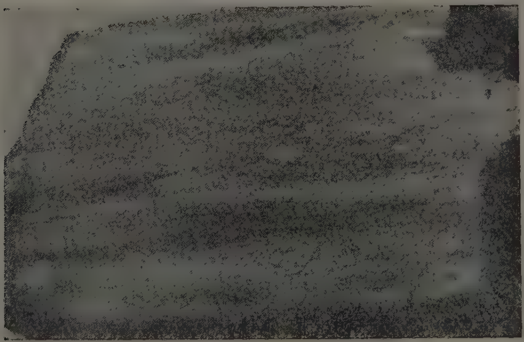

FIGURE 5. Parallel banding in copper pyrite ores replacing schistose siliceous shales. Urup deposit.

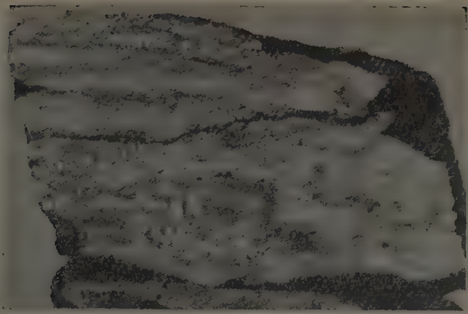

FIGURE 6. Curved banding of dispersed ores, deposited along slaty partings in quartz albitophyre. Urup deposit.

Thus at the initial formation stages of the ore body, and as a result of the formation of a fractured zone along the contact of lithologically different rocks, minerals of early mineralization phases, i.e., chiefly copper pyrite, mainly replaced the quartz albitophyres and inherited their texture.

Fine-grained pyrite in the pyritized quartz albitophyres occurs chiefly in slaty partings (Fig. 6) whereas its less common coarser crystals usually occur within the schistosity lenses.

The high concentration and the small size of pyrite crystals in the slaty partings of quartz albitophyre suggest that solutions over-saturated with this component moved along these partings, with a rapid precipitation of pyrite from the hydrothermal solutions and with the formation of a large number of crystallization centers.

The low concentration of pyrite crystals within the schistosity lenses in quartz-albitophyre, and their larger size, suggest that the pyrite-carrying hydrothermal solutions were here less concentrated; accordingly, fewer crystallization centers were originated. The distribution of pyrite crystals, in their number and size, depending on the texture of enclosing rock, suggests that the hydrothermal

FIGURE 7. Curved banding in copper pyrite ores replacing schistose quartz albitophyres.

Lenticular segments of coarse-grained pyrite with quartz, surrounded by bands of fine-grained pyrite. Urup deposit.

solutions penetrated the schistosity lenses from the parting. Consequently, this indicates a replacement of schistose rocks by ores.

Coming nearer the massive ores, the pyritization is intensified while preserving the former texture, until finally the dispersed ores change to a series of conformable copper-pyrite veins and then to banded copper-pyrite ores. Iron-pyrite ores, which have replaced quartz albitophyre, have the same type of banding as the dispersed ores; i.e., lenticular segments of coarse-grained pyrite containing some nonore minerals are fringed by bands of fine-grained pyrite containing a small amount of nonore minerals (Fig. 7). Thus the textures of banded iron pyrite ores are mostly relicts of the schistose-rock textures. An exception is ores with textures complicated by intramineralization shifts during the ore deposition. In that event, a more uniform banding is locally observed, for both the dispersed and massive ores. The massive textures of copper-pyrite ores have been formed in the more complete replacement of rock by pyrite.

In the following mineralization phases, zones of open fractures originated chiefly in siliceous metashales of the hanging wall.

FIGURE 8. Relict textures of ore-replaced rocks.

Upper half -- replaced fine-grained tuff, with uniformly parallel schistosity; lower half -- quartz albitophyre, replaced by ore, with curved schistosity. Vlasinchikha deposit.

Hydrothermal solutions of later mineralization phases were more saturated with copper and zonc compounds. For this reason, the copper-zonc and copper-pyrite ores chiefly replaced the siliceous metashales of the hanging wall, although preserving their texture. Intramineralization shifts complicated the textures of the nascent ores. As a result, there appeared veins of ore minerals occurring conformably in all varieties of ores. Rich copper ores in the hanging wall have originated in about the same way. The fairly consistent character of typical ore texture, both along the strike and the dip, together with their abrupt change with the thickness of the ore body, also confirms their relict origin. Furthermore, the most intensive fracturing of ores occurs in the central part of the ore body; going toward both the hanging wall and the floor, the degree of fracturing decreases. This also points at the absence of metamorphism, because otherwise the more intensive fracturing would have occurred along the periphery of the ore body, at any position of its varieties (Fig. 8).

Thus, the structures of the Urup ores are relicts of those prevailing in the replaced enclosing schistose rocks; in addition, they have been complicated by the subsequent mineralization phases.

STRUCTURES OF ORES AND MINERALS AS A RESULT OF THEIR ORIGINAL DEPOSITION

Changes in the composition of mineralizing solutions, with time, have determined the change in ore texture, in both sides of the ore body. Ores in the floor are made up of a coarser pyrite than in the hanging-wall ores.

The fine-grained texture of banded ore in the hanging wall has been determined not by the breaking-up of the grains in metamorphism but by a change in the crystallization conditions for pyrite, at different stages of mineralization.

The colloform grains of pyrite, in places, display a zonation wherein the central colloform part is surrounded by a concentric crystalline band. In places, the situation is reversed, with a colloform periphery and a crystalline center. Such pyrite-grain texture cannot be taken for evidence of crystallization of colloform pyrite in metamorphism, because in such an event, the crystallization should have proceeded from periphery to center.

The postmetamorphic formation of dispersed pyrite ores is obvious from the relationship between pyrite crystals and schistosity of the

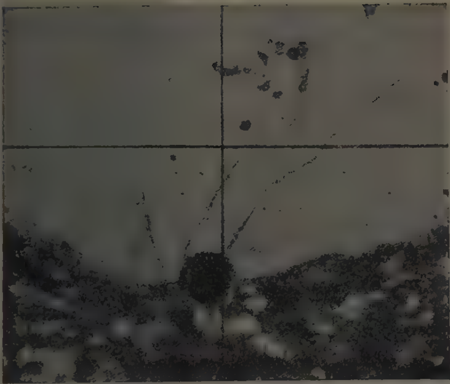

FIGURE 9. Deformation of porphyritic dispersed quartz by a pyrite crystal. Crossed Nicols. Urup deposit.

enclosing matrix. In crystallization, pyrite exercises considerable pressure on the latter and deforms it, with the crystals themselves commonly remaining intact (Fig. 9).

The mechanical action of a growing pyrite crystal on the surrounding matrix is demonstrated by the widening of schistosity partings with the accompanying typical association of minerals in the "pressure shadows" (Fig. 10). These crystals, and the "pressure shadows" with them, are postmetamorphic. Otherwise, they would have been deformed by the slaty partings.

A vast majority of the pyrite crystals in dispersed ores are euhedral, without any evidence of crushing. Only a small portion of them is broken, in ore making, by intramineralization shifts.

A constant companion of ore minerals in copper-pyrite ores is metasomatic quartz in irregular and serrate grains. It accompanies the pyrite aggregates and cements them in crushed segments. Such quartz is partially replaced by chalcopyrite and sphalerite. In most cases, its grains do not display the wavy extinction characteristic of metamorphosed quartz. It follows that this serrate quartz, precipitated from hydrothermal solutions before chalcopyrite and sphalerite, but after pyrite, has not undergone the regional metamorphism.

It follows, further, that ore minerals which partially replace the serrate quartz — such as sphalerite and chalcopyrite — have also not been metamorphosed.

THE SO-CALLED "PEBBLES" OF PYRITE ORES IN ROCKS OF THE HANGING WALL AND FLOOR OF THE ORE BODY

Metamorphic rocks in both sides of the Urup ore body in places carry lenselike bodies of pyrite ore, reminiscent of pebbles in their form and size. Such lenses occur in conjunction with pyrite veins, commonly in the same place, and are of about the same composition as the veins.

These pyrite lenses and veins are accompanied by calcite veins which fringe them but never intersect. There are pyrite lenses where the enclosing rock is incompletely replaced and

FIGURE 10. The lack of deformation of pyrite crystals by slaty partings in quartz albitophyre. Parallel Nicols. Urup deposit.

penetrates the lense in zig-zags. In some instances, the lenses carry relicts of unreplaced feldspar from the enclosing tuffs. In some places, they carry calcite in their central part.

Thus the genetic relation of these pyrite lenses with veins of pyrite and calcite — both undoubtedly hydrothermal formations — as well as the presence of relicts of the incompletely replaced tuff, and the presence of the lenses themselves in rocks both overlying and underlying the ore body; i.e., in rocks older than the latter — all show that these lenses are metasomatic formations rather than redeposited ore pebbles.

THE FORMATION OF PYRITIZATION ZONES, SUBSEQUENT TO THE SCHISTOSITY OF ENCLOSING ROCKS, AND THEIR GENETIC RELATIONSHIP WITH PYRITE ORE DEPOSITS

Noncommercial pyrite ore deposits, the so-called "pyritization zones," present fairly extensive and thick zones of vein mineralization and of dispersed pyrite ores, with a noncommercial content of useful components. These pyritization zones occur unconformably to the stratification and, in places, to the schistosity of the enclosed rocks, and are represented by a system of mineralized normal-type faults.

Veins of pyrite and chalcopyrite in these pyritization zones are either conformable with the schistosity or else cut across it. In their material composition and occurrence, they are similar to veins occurring in the vicinity of ore bodies; i.e., they are represented by normal fault planes filled with ore minerals and having a massive structure.

The presence of ore veins and mineralized fault planes which cut the schistose rocks, together with the greater intensity of disturbance within the pyritization zone, confirms its formation in already schistose rocks. Quartzitized quartz-albitophyres, cut by pyrite veins, locally change to secondary quartzites.

The presence of quartzitized schistose fragments of quartz albitophyres shows that the quartzitization affected already schistose rocks. It follows that pyrite veins which cut the quartzitized rock, too, are younger than the schistosity.

Quartzitized breccias of diabase porphyries, cut by pyrite veins, have been observed. The schistosity in the breccia fragments is irregularly oriented, which indicates a crushing of the already schistose rocks. The quartzitized breccia zone is cut by veins of iron pyrite, which follow joint planes. These veins are lenticular, parallel to each other, with no evidence of crushing. Consequently, these pyrite veins, too, are postmetamorphic.

Besides the iron-pyrite veins, there are veins of sphalerite which cut both the schistose rocks and the massive secondary quartzites.

It appears then, from the relationship of ore veins in the pyritization zone with the schistosity of enclosing rocks, and from their younger age as compared with the brecciation of the schistose rocks, that the pyritization zones, along with the ore bodies, are younger than the schistosity and have not undergone regional metamorphism.

Ore bodies of commercial pyrite deposits change to noncommercial zones of pyritization, both along the strike and along the dip. These zones are distributed among the (commercial) pyrite deposits, connecting the latter both genetically and spatially. Accordingly, the entire mineralized complex of the Urup pyrite-deposit area can be regarded as a contemporaneous postmetamorphic formation.

CONCLUSIONS

Data on the relationship of ore bodies, zones of pyritization, and their textures, with the schistosity of enclosing rocks and dikes which cut the latter, lead to the conclusion that ore bodies in the Urup pyrite deposits were formed in already schistose rocks and have not undergone regional metamorphism.

The lower age boundary for the Urup pyrite deposits is the time of intrusion of the preore dikes which cut the Devonian metamorphics; the lower boundary is middle Carboniferous deposits not affected in this area by either the igneous or hydrothermal activity.

The age of the metamorphic sequence, which we have assigned to the Devonian, is uncertain. An older age is not to be ruled out.

The Urup ore bodies and zones of pyritization represent mineralized zones of postfolding normal faults.

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BRIEF COMMUNICATIONS

METASTABLE K-FELDSPAR AND ZEOLITE IN ORES OF THE DAL'NETAYEZHNYY (FAR TAIGA) DEPOSIT[1]

by

V.N. Vasil'kova

A specific feature of the Primor'ye (Maritime region) sulfide-cassiterite deposits, as noted by all students beginning with S.S. Smirnov, is their formation under near surface conditions, a considerable distance away from any magmatic hearth. The recent work of the tin group, All-Union Institute of Mineral Resources (VIMS) has established the close relationship between the formation conditions for cassiterite-sulfide deposits and the depth of the intrusions. Depending on the depth, the following deposits have been recognized:

1. closely related, both spatially and genetically, to magmatic bodies now exposed on the surface,

2. those revealing no direct connection with magmatic sources but exhibiting contact-metasomatic halos, with a constant change in the mineralization, going away from the sources,

3. those displaying neither direct nor indirect connection with intrusive rocks.

For deposits of the first two groups, the combination of near-surface conditions of formation and the relative proximity of the sources of mineralizing solutions indicates a high initial temperature of the ore crystallization on one hand, and a rapid drop in it in the process of mineralization on the other. This is reflected in the appearance of very unbalanced mineral forms and of peculiar mineral associations. An interesting instance is the Dal'netayezhnoye (Far Taiga) tin ore deposit in Primor'ye.

This deposit is represented by a complex system of mineralized zones of crushing in a Mesozoic arenaceous and argillaceous sequence. Igneous formations of the area consist of thick dikes of granite porphyries, and assorted porphyrites and lamprophyres. They are all preore and commonly control the mineralization.

The relationship between granite porphyries and basic dikes is readily established from the fact that the latter cut the former. No other igneous rocks have been discovered in the ore-field area. However, a well-defined halo of intensively biotized rocks occurs in its southern part. Going north from it, quartz-cassiterite ores change to chiefly sulfide.

The formation of ores proceeded in stages, which determined a broad development of brecciated and breccialike ore textures. The following main mineralization stages have been identified: 1) arsenopyrite-cassiterite-quartz with K-feldspar; 2) pyrrhotite; 3) pyrite-ankerite; 4) quartz-adularia, culminating in the deposition of later pyrite and calcite.

Metasomatic replacement is the predominant deposition type within the crushing zone, and filling of fractures also takes place.

We shall describe only the most interesting ore minerals, K-feldspar and zeolite, whose presence in association with cassiterite sheds the most light on the conditions of ore deposition. The cassiterite-zeolite association is interesting also because it has not been previously discussed in the literature.

Two generations of K-feldspar are clearly defined in ores of this deposit. The first is associated with quartz-cassiterite ores; the second has been noted in an intergrowth with later and barren quartz.

The first generation K-feldspar is precipitated in euhedral rhombic crystals and xenomorphic grains, from less than 1 to 3 mm, subordinate in their form to quartz and cassiterite. Its segregations in isolated veins attain several centimeters across. In places, K-feldspar replaces quartz, to form an intricate corrosion structure reminiscent of the micrographic. Its later precipitation, with relation to quartz, is indicated by the

[1] Metasabil'nyy kaliyevyy polevoy shpat i tseolit v rudakh dal'netayezhnogo mestorozhdeniya.

fact that quartz-sericite veins are cut by the quartz-feldspar.

The precipitation of K-feldspar is associated with the end of the quartz-cassiterite stage; however, K-feldspar continued to be deposited in the beginning of the sulfide stage. This explains its frequent presence in quartz-sulfide ores where it usually occurs near fahlbands of their plumate fractures, in association with quartz, sulfides, carbonates, and some scheelite.

In all instances, it is developed chiefly in the metasomatic ore segments, growing together with quartz, in fringes of fragments of quartzitized rocks, less commonly directly within the vein, in the interstices between quartz grains.

In massive accumulations, K-feldspar is milky white; in thin fragments it is water-transparent. Its luster is vitreous. In thin sections, it is observed both in regular rhombic crystals and in irregular grains without cleavage. In rare instances, it is definitely cataclastic. The refraction index, measured in immersion preparates, is α = 1.517 ± 2; γ = 1.522 ± 2. Most interesting is the fluctuation of the optical-axes angle. As measured on the Fedorov table, this angle varies from 0° to 80°, for different grains, even in the same section. These fluctuations occur in both the rhombic sections of crystals and in the irregular grains. Out of 21 measurements in 5 thin sections, one-third of the grains had the optical axis angle of 0 – 10°; in four instances, -2V varied from -60° to -80°; for the balance, it had an intermediate value of -36° to -60° (Fig. 1). Similar fluctuations have been established within a single grain: in the central part, -2V = 0°; for the periphery, -2V = -56°.

An x-ray analysis by G. A. Sidorenko

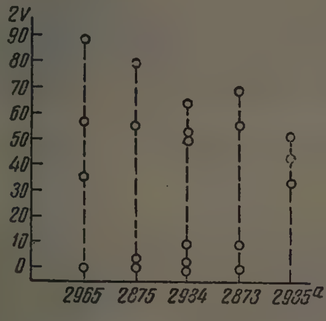

FIGURE 1. Change in angle 2V, in early generation K-feldspar.

revealed a monoclinal type of lattice, peculiar to K-feldspar with an irregular distribution pattern of aluminum and silicon atoma.

The second generation K-feldspar, too, is developed in euhedral rhombic crystals, in a matrix of barren hornfelslike quartz with a relict concentrically-zonal structure; or else it forms a rhythmic alternation with columnar chalcedonylike quartz and with zones of cryptocrystalline quartz with a festoon texture.

The two feldspars are identical in their physical and optical properties, as established by x-ray analysis and by measurement of the refraction index. The difference is only in their angles of optical axes. As seen in Fig. 2, the second generation K-feldspar is characterized by an optical axes angle of 50-90°, with only isolated measurements revealing a value of -2V = -30 – 40°.

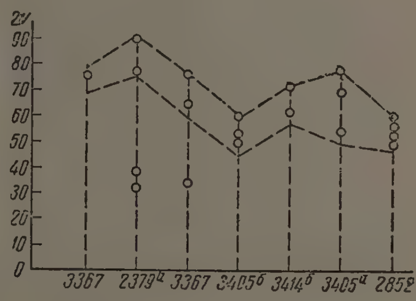

FIGURE 2. Change in angle 2V in late adularias.

These investigations make it possible to assign K-feldspar to a metastable sanidine-like form which changes in the process of ore crystallization to adularia, typical of hydrothermal deposits.

Sanidine is known to be the highest temperature variety of orthoclase, typical of porphyritic lava separates and of some extrusives. The possibility of existence of hydrothermal sanidine was pointed out by Academician D. S. Belyankin. In 1936, K. N. Fenner described low-temperature sanidinelike feldspars with 2V = -33 – 51°, from the Yellowstone geyser area, U. S. A., formed at 100 – 200° C. The crystallization conditions for such unbalanced forms apparently do not differ on the whole from those for other metastable minerals, such as tridymite and cristobalite, also originating at temperatures considerably differing from the limit of their true stability.

Calcitic zeolite-leonhardite (laumontite) is the main component of cassiterite-zeolite ores which make up one vein in the same deposit. Isolated pockets of similar ores occur also in sandstones near the Raduzhnaya vein.

In both instances, the ores were formed as a result of the single filling of fractures and pockets. The zeolite of these ores is closely associated with quartz, cassiterite, arsenopyrite, and an assemblage of Ca-minerals: actinolite, fluorite, axinite, and epidote. It is xenomorphic in relation to all these minerals, being precipitated in inter-

Table 1

Interplanar spacings in zeolite (specimen 2991-56)

1	6,948	8	23	1,952	5
2	6,251	3	24	1,861	4 scheelite (?)
3	5,067	3	25	1,795	2
4	4,503	5	26	1,861	3
5	4,142	10	27	1,703	1
6	3,862	2	28	1,626	7
7	3,665	3	29	1,592	2
8	3,508	10	30	1,567	2
9	3,385	3	31	1,523	5
10	3,270	3	32	1.480	3
11	3,198	2	33	1,437	6
12	3,016	6	34	1,344	3
13	2,855	5	35	1,303	5
14	2,787	4	36	1,262	6
15	2,570	3 scheelite (?)	37	1,232	5
16	2,445	6	38	1,194	4
17	2,368	4	39	1,166	3
18	2,280	2	40	1,150	3
19	2,214	2	41	1 119	2
20	2,152	7	42	1,090	4
21	2,094	2 scheelite (?)	43	1,047	3
22	1,995	1	44	1,024	3

Table 2

Chemical analysis of zeolite

Oxides	Weight %	Molecular quantity	Atomic amount of oxygen	Atomic amount of cations	Number of cation atoms
SiO_2	54.78	912	1824	912	17.5
Al_2O_3	21.25	208	624	416	8.0
Fe_2O_3	0.52	3	9	6	0.1
CaO	10.72	191	191	191	3.67
MgO	0.80	20	20	20	0.38
K_2O	0.23	2	2	4	0.08
Na_2O	0.36	6	6	12	0.23
H_2O	11.65	653	653		12.5
Total	100.31		3329		

From this analysis, the following formula has been derived for zeolite, on the basis of 48 atoms of oxygen:

$$Ca_{3.67}(Na_{0.08}\cdot K_{0.23})Al_8Si_{17.5}O_{48}\cdot 12.5\ H_2O$$

which is very close to the ideal formula of laumontite:

$$Ca_4Al_8Si_{16}O_{48}\cdot 16H_2O.$$

stices among the grains, or else enclosing the latter in poikilitic inclusions. The richest ore varieties are made up of abundant aggregates of fine-grained cassiterite in massive zeolite.

The standard subhedral texture of the ores, the absence of zeolite crossings of all of these minerals, as well as the absence of any evidence of replacement for earlier minerals, precludes the possibility of zeolite precipitation at a later mineralization stage.

Macroscopically, zeōlite is recognized by its brittle fibers, which are white with a vitreous luster. There is a well-defined cleavage along two planes at an angle of 51°. Elongation positive; $c\gamma$ = 40°; 2V = -32°; refraction indices, γ = 1.516; α = 1.506 ± 2. It is readily dehydrated in heating, with liberation of much water.

Given above are the results of x-ray study by G. A. Sidorenko, and of chemical analysis by L. Polupanova (Tables 1 and 2).

Some excess silica apparently is the result of the mechanical addition of quartz. The deficiency in water makes it possible to assign the mineral to a slightly dehydrated variety of laumontite, designated as β-leonhardite by A. Ye. Fersman.

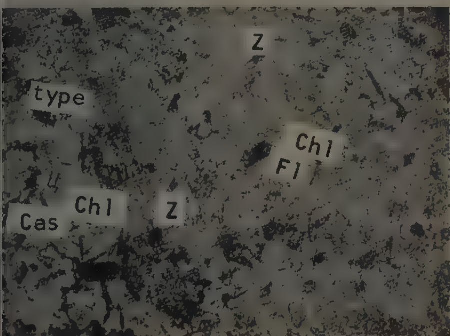

FIGURE 3. Zeolite (Z) in association with fluorite (Fl), chlorite (Chl), and cassiterite (Cas).

Leonhardite has been described from copper-zeolite ore deposits at Lake Michigan, U. S. A., and from a number of large deposits in Romania. However, it is best developed in cavities of extrusive rocks. For comparison, we present the chemical composition and optical constants for leonhardite which S. D. Coombs studied in detail among the replacement products of vitreous tuffs of New Zealand (Table 3).

Table 3

Chemical composition (% by weight)		Optical constants
SiO_2	— 52,04	α=1,507
TiO_2	— —	β=1,516
Al_2O_3	— 21,46	γ=1,518
Fe_2O_3	— 0,12	—2v=26°±4°
MgO	— —	$\gamma\Delta c$=32°
CaO	— 11,41	
Na_2O	— 0,20	
K_2O	— 0,66	
H_2O	— 13,80	
Total	— 99,69	

The chemical composition for this leonhardite is expressed by the following formula, computed by S. D. Coombs on the basis of 48 oxygen atoms:

$$Ca_{3.79}(Na_{0.12}K_{0.26})Al_{7.96}Si_{16.04}\cdot O_{48}\cdot 16H_2O.$$

As seen from a comparison of their optical properties and chemical composition, both minerals are very similar.

The association of zeolite with cassiterite, actinolite, axinite, arsenopyrite, as well as the above-described metastable sanidinelike varieties of K-feldspar in ores, are evidence of an abrupt change in temperature during the mineralization process.

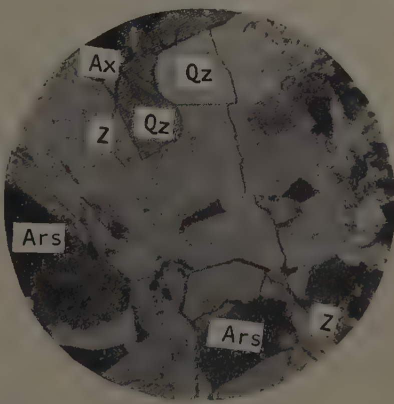

FIGURE 4. Subhedral texture of zeolite ore.

Z - zeolite; Ax - axinite; Qz - quartz. Thin section 2484; X-36, with analyzer.

The temperature change is explained by the nature of the deposit, apparently formed a short distance away from a deep magmatic hearth, on one hand, and approaching the surface in its upper part, on the other.

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REVIEWS AND DISCUSSIONS

I. YA. DANILANS' BOOK, "HOLOCENE FRESH-WATER CALCAREOUS DEPOSITS OF LATVIA"[1, 2]

by

N. N. Sokolov

This work of I. Ya. Danilans fills a sub-tantial gap in lithological literature. It pre-ents an excellent description of recent fresh-'ater carbonate deposits[3] and, what is espe-ially important, suggests and partially stablishes the patterns in their origin and istribution throughout Latvia. Credit is given) previous investigations, which have been rogressing since Grevingk (1861), and espe-ially for the last 30 years, in connection with ie calcification of soil.

Up to now, the published works in this field ave been chiefly of a narrow empirical char-cter, without a proper substantiation of the rigin (of rocks) and without a demonstration f their main geographic relations and pat-rns. The formation and distribution of most ecent fresh-water carbonate deposits, how-ver, has been determined by a close and ommonly complex interaction of various geo-raphic factors, undergoing appreciable changes oth in space and time.

I. Ya. Danilans tackles a wide range of pics, from terminology and a historic re-ew, to the age of the deposits and to the leogeographic conditions of their formation. addition, the author has given a brief sum-ary (chiefly in tables) of numerous data — emical, granulometric, petrographic — and information on the reserves and exploitation of 50 large deposits. Of interest are data on pollen of woody plants in these deposits. The author cites many references to the work of his predecessors — in Latvian, and therefore little known — and takes into account material extant on the deposits in other parts of the Soviet Union and abroad. A sizable list of references is appended.

Obviously, I. Ya. Danilans' monograph is very useful. It will promote a more comprehensive knowledge of most recent carbonate deposits, not only in Latvia, but elsewhere in this country.

As stated in the text, there are about 900 known and studied deposits in Latvia, with an overall reserve of 23 million cubic meters; among them, 25% are large (10,000 to 200,000 m^3) and 2% very large (more than 200,000 m^3). The author notes that most deposits have been formed by the leaching of carbonates from Quaternary beds.

On the basis of geographic conditions, he identifies the following "morphologic types of calcareous spring deposits:" 1) cliffs; 2) slopes; 3) foothills; 4) domal deposits; and 5) deposits in topographic lows. Domal deposits are formed by pressure (artesian) waters and are located in the vicinity of foothills; in their character they are close to the foothill-type deposits.

In addition, two morphologic types of lacustrine deposits are designated: 1) shallow basins and 2) deeper lacustrine troughs.

Foothill deposits of the slopes of valleys, ravines, and hills are best developed among the spring deposits; shallow basin deposits predominate among the lacustrine. The former are comparatively small; the latter commonly grow very large. Comparatively rare but large deposits are associated with topographic lows and with deep lacustrine troughs.

The deposits themselves are described by the author (p. 40) as "friable to consolidated

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[1]O knige I. Ya. Danilansa "golotsenovyye presnovod-'ye izvestkovyye otlozheniya latvii.

[2]Published by the Academy of Sciences, Latvian S.R., Riga, 1957, p. 151.

[3]The latter are widespread in Latvia where they are great practical importance.

accumulations of aggregates and isolated crystals of calcite, of various size and form, with a varied amount of other minerals and organic matter." Microscopically (the book is illustrated with photomicrographs), there are two main types of deposits: 1) friable and 2) consolidated. The first type is subdivided into mealy and granular varieties; the second is subdivided into lumpy, slightly consolidated, and well consolidated. Consolidated deposits account for only about 5 percent of the total; the most common are unconsolidated deposits of small calcite aggregates, usually cryptogranular and pelitomorphic.

The author dwells in detail on impurities usually present in the deposits, with their quantitative and qualitative composition substantially differing even within a single deposit. Three groups of impurities are recognized: clastic (arenaceous to mealy argillaceous), organic (peat and organic oozes), and ferruginous (limonite). There are small amounts of $MgCO_3$ (as much as 2%) and P_2O_5 (not more than 0.1%).

Depending on the amount and character of the impurities, the author proposes two types of lime recognizable in the field; 1) fresh-water and 2) contaminated fresh-water.

The first type includes deposits without any definite impurities; the second contains argillaceous, sandy, ochre-colored, peat, and ooze impurities. The minimum content of $CaCO_3$ in deposits of the second type varies from 30 to 70 percent, depending on the character and amount of impurities.

In dealing with the age of the deposits, the author believes, on the basis of the composition of woody plants' pollen, that most of them were formed during Boreal time. Of interest are his views on the conditions of lime formation in fresh waters, and on the distribution of the deposits throughout Latvia. However, definite patterns on the distribution of these deposits could not be worked out.

The basic premises and conclusions of the author are of obvious scientific and practical importance. However, his work calls for a few remarks.

1. We believe it would be more correct to call the deposits, carbonate, rather than calcareous.

2. It would be well to cite more instances of deposits in different geomorphologic and landscape environments.

3. As long as the features of relief and of Quaternary deposition have been correctly assigned a major part in the formation of deposits, they should have been studied in more detail, be it only within the deposit areas.

4. The same is true for the study of soils, inasmuch as soil processes play an important part in the formation of most recent carbonate deposits. Specifically, the Boreal age of the deposits can be explained on the basis of the evolution of soils. It is quite probable that intensive leaching of carbonates from soils became possible only in Boreal time. Before that, it was evidently hampered by permafrost. Later on, after the Boreal leaching of soils, carbonates could be flushed out only in rare instances, chiefly from surface outcrops.

It should be possible to organize a comprehensive study, if only in a few typical area, with participation of geomorphologists and pedologists.

5. As is frequently noted by the author, the distribution of deposits is closely related to topography. Consequently, the regularity in their distribution is better understood first of all through an analysis of geomorphologic data. It seems to us that the areal differentiation of the country (especially for the better-known western part of Latvia.

This well-written and excellently published work of I. Ya. Danilans deserves the highest praise. It also should be noted that, together with the earlier publications (N.A. Ansberg, E.B. Rinks, Ya.Ya. Selitskaya; Most Important Quaternary Clays of the Latvian S.S.R., 1955; Useful Minerals of the Latvian S.S.R. 1. Carbonate Rocks, 1957), it bears testimony to the quality of lithologic study in the Geology Institute, Academy of Sciences, Latvian S.S.R.

THE STRATIGRAPHIC OUTLINE OF TRIASSIC CONTINENTAL DEPOSITS IN EASTERN URALS[4]

by

T.A. Sikstel'

Issue no. 5, vol. XLIII, 1958, of "Botanical Journal," carries a short article by A.I. Turutanova-Ketova, "Floral Characteristic of Some Lower Mesozoic Productive Intervals on the Eastern Slope of the Middle Urals." This article is of more interest to geologists than to botanists, because it gives more space to stratigraphy and paleogeography than to the description of fossil plants.

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[4]O stratigraficheskoy skheme dlya kontinental'nykh otlozheniy triasa vostochnogo urala.

The author studied assemblages of plant remains collected from Triassic deposits of two areas in the eastern slope of the Urals. Triassic continental deposits are very little known, on the whole; this is especially true for their stratigraphy. The work of A.I. Turutanova-Ketova is significant, in this respect. However, a careful perusal of the "Botanical Journal" article calls for caution regarding the author's conclusions.

The article gives a paleogeographic outline of the Triassic for the area in question, based on the author's stratigraphic classification. The classification is especially interesting: the author breaks up the continental Triassic into divisions; and she breaks up the upper division into stages. All of the material is presented in three tables. The first two contain a lithologic description of the formations, and their thickness. There are special columns for "index plants," "data of spore-pollen analysis, fauna," and "age." Column "data of spore-pollen analysis, fauna" is empty in five instances out of eight. In one instance (Bulanash formation, Noric stage), it says, "Fresh-water pelecypoda of Triassic aspect; less common, transitional to Lower Jurassic;" in another (motley formation, Middle Triassic), "Pollen and spores," without mentioning the composition; again (middle formation, Middle Triassic), "Spores and pollen, similar to Early and Middle Triassic." It follows that neither the fauna nor the data of spore-pollen analysis was the basis of stratigraphic differentiation for continental deposits; accordingly, the latter must have been based on the column "index plants." The author identifies a total of 46 forms, of which 13 are "index." It is nowhere made clear, however, what is being understood as an "index form." Therefore, the presentation of material and its significance are not convincing.

The Bulanash-Yelkin depression (Upper Triassic (Table 1) is subdivided into Karnic, Noric, and Rhaetic. However, the flora rosters in column "index forms" reveal that all forms present in the Rhaetic (Bobrovsk formation) are also present in the Noric (Bulanash formation). The situation is almost the same in the correlation of the plant assemblage with the Karnic stage (Yelkin formation). We quote from the text, the better to explain the author's views.

A reason for the stratigraphic differentiation is given on page 669: "The oldest motley formation . . . has been tentatively dated back to the Ladinic, Middle Triassic. It is characterized by a single form, *Pityolepis cedriformis* sp. nov. This isolated seed scale is coniferous, probably belonging to family Pinaceae; morphologically, it is very close to those of *Cedrus*. V.V. Zauer dates the appearance of cedar back to Late Permian (by microspores)." Why, then, does A.I. Turutanova-Ketova assign the motley formation, be it only tentatively, to the Ladinic rather than to Anisic, Karnic, or Tatarian?

The motley formation rests upon the weathering crust of a volcanic sequence which is also conditionally assigned to the Lower Triassic. It is overlain by another weathering crust, so that there is no reason to assign it, however conditionally, to the Ladinic even on the basis of its position. Further on, in substantiating the age of the Telkin formation, presumably Karnic, the author mentions the scarcity of index forms — only 6 typical of the Chelyabinsk province Rhaetic, out of the assemblage of 23 species; also the fact that two forms, *Taeniopteris ensis* and *Juccites* sp., are known from the Surakay Lower and Middle Triassic. Unfortunately, the author does not mention that these forms are known not only from the Tananyk (Krivlevskaya) formation of Middle-Upper Triassic, but also from the overlying Upper Triassic Surakay formations. In the following exposition, the author concludes that it is impossible to determine the age of the Yelkin formation from the plant assemblage. "Therefore the conclusions as to the age of deposits carrying these remains must be based on the stratigraphic relationships of the consecutively deposited formations, and the Yelkin formation must be assumed to belong to early Keuper; i.e., to the Karnic." It is incomprehensible as to why the author specified the Karnic.

In the Bulanash formation, a plant assemblage of 26 forms was identified by A.I. Turutanova-Ketova. There are some errors in the treatment of the vertical distribution of some of the species. For instance, *Lobatannularia* is known in Fergana not only from Upper Permian and Lower Triassic deposits but from the Liassic as well [1]; whereas *Danaeopsis* from Madygen (Fergana) occurs in early Keuper and not in the Lower Triassic [3].

The author regards the *Juccites* group (p. 673) as very important in the age determination. According to her, this group persisted from the beginning to the end of the Triassic. She concludes, "The *Juccites* group, in conjunction with other index plants, dates the Bulanash formation as upper Keuper Noric." But why the Noric? The fact is that *Juccites* persists from Lower to Upper Triassic; *Schizoneura* — from Middle Triassic to lower Liassic; *Lobatannularia* — from Permian to Liassic; *Danaeopsis* — from Lower to Upper Triassic. Consequently, plant remains do not give a basis for such a differentiation. Equally controversial is the statement that *Najadites* sp. is "a definitely Triassic form." This genus is known from the Carboniferous on. The Bobrovsk formation is tentatively designated as Rhaetic, from the presence of *Schizoneura* sp.

and Juccites sp. "which marks the upper limit of their Late Triassic distribution." This is despite the fact that the presence of Juccites was used as a proof of a Noric age of the Bulanash formation. The problem of stratigraphic differentiation of the Anokhinskaya depression Triassic is solved by the author in the same way. A barren sequence at the base of the section is tentatively assigned to the Lower Triassic, "by virtue of its position in the section." The middle formation is assigned to the Middle Triassic because of the presence of Lepidopteris ottonis, Sphenòcallipteris uralica (new species), and Stenopteris cf. elongata Carr.; on the subject of Lepidopteris and Stenopteris, the author states, however, that they are typical of the Rhaetic of Greenland, Sweden, and other places. To be sure, there is a statement to the effect that representatives of these genera are known from the Surakay Lower-Middle Triassic. But is that a proof of the presence of specifically Middle Triassic in the Urals?

It appears that the article is not adequately edited. For instance, in Table 3, Taeniopsis ensis is given as an index form for the Yelkin formation alone, whereas in Table 1 it figures for both the Yelkin and Bulanash formations. In Table 1, Juccites is stated for two formations; in Table 3, for three, with a symbol for the index form. We read on page 669, "A characteristic form in this sequence (Yelkin formation; T.S.) is Taeniopteris ensis Oldh. and Juccites sp." In Table 3, however, Juccites sp. is altogether missing, yet the Yelkin formation is said to carry Juccites uralensis Pryn., present in both the Bulanash and Bobrovsk formations.

Material at the disposal of A.I.' Turutanova-Ketova is obviously inadequate for a detailed (substage) stratigraphic scheme. Paracalamites sp., Schizoneura sp., Lobatannularia sp., Neocalamites sp., and Lepeophyllum (?) sp., given by the author as index forms, may serve as a basis for differentiation of the enclosing sequence from those older than Upper Permian and younger than Jurassic. They do not warrant, however, the assumption of a Karnic, Noric, or Middle Triassic age for deposits containing these fossils. The recurrence of the same forms in formations of the Bulanash-Yelkin depression does not consititute a paleontological proof of their age.

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LETTER TO THE EDITORS OF IZVESTIYA OF THE ACADEMY OF SCIENCES, U.S.S.R., GEOLOGIC SERIES

In connection with the publication of G.I. Teodorovich's letter in no. 7, 1958 of your magazine, I request the publication of the following statement of fact:

1. I have never attributed any oil generating properties to extrusive rocks, either basic or acid. I have associated, and do associate now, the origin of oil with hydrocarbon radicals liberated from deep-seated magmatic hearths. (Materials of the Discussion on the Origin and Migration of Oil, Kiev, 1955, pp. 73, 74, 76), rather than with extrusive rocks, formed out of the magma. The statement of G.I. Teodorovich, that "according to N.A. Kudryavtsev, only the basic extrusives are oil generating," is not true; magma and extrusive rocks are far from being synonymous.

2. Oils shows, associated with Fennoscandian extrusives, are very common and diversified (G.I. Teodorovich inquires in his letter as to why they are not known). There are gas shows (Hellivara, Khibiny, etc.) and oils seeps. (Dannemora, Grengesberg, etc.); whereas the solid bitumens of Sweden occur locally in such quantities in iron veins, as to decrease the expenditure of coal in the smelting of the latter, by 30 to 40 percent. Numerous oil shows in the Archaean of Fennoscandia are mentioned in manuals of Egler-Heffer (1909) and Redwood (1926). Some information on them is contained in texts by I.M. Gubkin and I.M. Gubkin and S.P. Kiselev; also in K.I. Bogdanovich's text, "Ore Deposits," vol. 1 (1912). M.F. Mirchink and A.A. Bakirov (Collection "Origin of Oil," pp. 424-425) are the only ones, as far as I know, who deny the occurrence of oil in the shields.

3. G.I. Teodorovich's paper in the Izvestiya of the Academy of Sciences, U.S.S.R., Geologic Series, no. 8, 1956, carries the following statement, after having quoted my paper

on the occurrence of oil shows in basement gneisses of the Volga-Ural oil province: "A.M. Mel'nikov and other geologists of the Trust categorically deny these 'facts' cited by N.A. Kudryavtsev." This is contrary to the statement of G.I. Teodorovich himself, to the effect that he denied only the "seeping of oil along fractures" from the basement, rather than "the oils shows themselves." In his paper, there was no question of the treatment of the facts but of the reality of these facts.

I am refraining from comment on other points of G.I. Teodorovich's letter, because this would take me beyond the statement of fact.

November 12, 1958 N.A. Kudryavtsev

CHRONICLE

ACADEMICIAN A.YE. FERSMAN MEMORIAL (1883-1958)[1]

This capital city took notice of the 75th anniversary of Academician Alexander Yevgen'yevich Fersman, an outstanding scientist of this country, who passed away on May 20, 1945.

A meeting dedicated to the memory of Alexander Yevgen'yevich [Fersman] was held on December 23, 1958, in the conference hall of the Presidium of the Academy of Sciences, U.S.S.R. It was organized by the divisions of geologic-geographic and chemical sciences of the Academy; by the A.Ye. Fersman Mineralogical Museum, Academy of Sciences, U.S.S.R.; by the Institutes of the Academy: Geochemistry of Ore Deposits, Petrography, Mineralogy, and Geochemistry; Mineralogy, Geochemistry, and Crystallography of Rare Elements; the V.I. Vernadskiy Institute of Geochemistry and Analytical Chemistry; and by the All-Union Mineralogical Society.

In the opening tribute, Academician D.I. Shcherbakov, Academic Secretary of the Division of Geologic-Geographic Sciences, Academy of Sciences, U.S.S.R., commented on the fruitful scientific activity of A.Ye. Fersman as one of the founders of the new science of geochemistry, his extensive study of the natural resources of this country, specifically of apatites in the Kola Peninsula and other northern areas, sulfur deposits of the Karakums; his predictions of the mineral wealth of Cheleken Island; his numerous trips to unexplored and barely accessible regions of the Soviet Union and abroad, the data of which he used for his interesting writings as a geographer-explorer; his achievements as a scientist-organizer, especially during the Great Patriotic War of 1941-1945, when he initiated and directed the Commission of the Academy of Sciences, U.S.S.R., for the geologic-geographic service of the Red Army. Academician D.I. Shcherbakov especially emphasized the work of A.Ye. Fersman as a talented popularizer of science, who has reflected in his writings the "poetry and life" of stones.

The pupils and followers of A.Ye. Fersman spoke on diverse aspects of the many-sided scientific work of this outstanding figure in the history of our national science.

Corresponding Member of the Academy, A.A. Saukov, read a paper on A.Ye. Fersman as a Geochemist.

He stated that geochemical work occupies an outstanding place in the rich heritage of A.Ye. Fersman. Together with V.I. Vernadskiy, he is rightly said to have been a creator of modern geochemistry. To him belongs the most comprehensive and up-to-date definition of geochemistry as a science of the distribution and behavior of chemical elements under various physical chemical conditions prevailing in nature. A.Ye. Fersman deserves especial credit for his work in the solution of the frequency problem for elements, throughout the earth. On the basis of numerous data on the distribution of elements in rocks, meteorites, water, atmosphere, and living organisms, he compiled many tables on the Clarke indexes for various geospheres. He analyzed the figures so obtained and explained them on the basis of the stability of atomic nuclei and of subsequent migration processes. He has shown that the most widely distributed, are those stable elements at the beginning of Mendeleyev's system; and among isotopes, those atoms whose mass numbers are the multiples of four.

A.Ye. Fersman attributed not only purely academic significance to the frequency of elements. He believed that it is destined to play an important practical part, inasmuch as the uneven distribution of elements is related to their commerical concentration in corresponding deposits.

He was much concerned with the migration of chemical elements, considering the problem

[1]Pamyati akademika A.Ye. Fersmana (1883-1958).

both for the properties of the migrant atoms themselves (internal factors of migration), and also against the background of the thermodynamic and physical chemical conditions of the medium (external factors of migration). From this point of view, he analyzed the behavior of various chemical elements under magmatic, pegmatitic, hydrothermal, and hypergenetic (surface and shallow destruction zone) conditions, paying special attention to the causes of the dispersion of elements and to their commercial concentration. He always emphasized that commercial deposits of useful minerals had been determined chiefly by the migration of atoms.

In later years, A. Ye. Fersman gave much effort to the substantiation and development of the geoenergy theory of migration. In this connection, he proposed a very simple general method of computation of the energy of crystalline lattices in minerals, on the basis of the energy coefficients for ions. He also introduced the concept of paragenes for ions and minerals, by which he meant the functions determining the time and place for the precipitation of minerals from cooling melts and solutions.

On the basis of a detailed analysis of the Mendeleyev system, A. Ye. Fersman worked out his well-known geochemical classification of elements. He explained their natural associations, on the basis of the isomorphism phenomena, and demonstrated the value of the system in the explanation of a number of other geochemical regularities.

Of great importance in the modern study of regularities in the distribution of useful minerals are his works on the regional geochemistry of Russia, the Kola Peninsula, Far East, and other regions, as well as his numerous investigations in the geochemistry of individual elements and deposits.

The theoretical work of A. Ye. Fersman was always closely tied to current practical problems. This was especially well demonstrated in his investigations in the Kola Peninsula and in Central Asia, as well as in the working out of the theory of mineralogic and geochemical prospecting methods.

Many present investigations by Soviet geochemists follow the path of their teacher.

Corresponding Member of the Academy, V.I. Smirnov, spoke on applied geology and geochemistry as reflected in the work of A. Ye. Fersman.

He related how the exactness of perception of a geochemist and mineralogist was combined with the energy necessary for the procurement of new raw mineral material for industry, in the active life of this scientist. The most momentous achievements of A. Ye. Fersman in this field, were associated with his many years in the Urals, Trans-Baikal region, Central Asia, and especially the Kola Trans-Polar regions.

A fundamental monograph by E. Ye. Fersman on pegmatites not only contains first-class scientific material summing up the results of his study for a quarter century but also generalizes the information on the practical significance of these peculiar geologic formations, a valuable source of rare metals and of a whole range of useful nonmetal minerals.

His four-volume "Geochemistry" has provided geologists with a tool which has sharply increased prospecting efficiency.

Many trends in modern prospecting for various groups of useful minerals are based on A. Ye. Fersman's work, and shall benefit, for a long time to come, from his fruitful ideas.

G.P. Barsanov, doctor of geologic and mineral sciences, described the scientific work of A. Ye. Fersman in the field of mineralogy.

A pupil of an outstanding scientist, Academician V.I. Vernadskiy, and a graduate of Moscow University, A. Ye. Fersman raised mineralogy to a new high level, in his charting of new paths in the understanding of chemical processes in the earth's crust, and in adapting to mineralogy the ideas and methods of chemistry, physical chemistry, and thermodynamics. A precise description of minerals and their occurrence, an analysis of the chemism of natural mineral-forming processes, and major generalizations in the history of the formation of mineral deposits in the framework of regional geologic history — such was the range of scientific interests and creative work of A. Ye. Fersman.

Among numerous problems in mineralogy, worked out by A. Ye. Fersman, that of pegmatites is the most fruitful in the development of genetic mineralogy. He worked on these peculiar mineral bodies from 1909 to his death. His efforts virtually created a new theory of pegmatites, their origin, classification principles, and the features of mineralogy and geochemistry of the pegmatitic process, in both granites and alkaline rocks. Practically, his work on pegmatites is closely connected with the discovery of a number of new pegmatite deposits carrying rare elements, precious stones, ceramic raw material, mica, etc. (Kola Peninsula, Karelia, Urals, Trans-Baikalia, and other places). We are indebted to A. Ye. Fersman for the "discovery" of pegmatites as

mineral bodies of a great practical value. Partly in connection with the mineralogy of pegmatites, A.Ye. Fersman devoted a great deal of attention to the study of precious stones. His work in the mineralogy of diamonds, his descriptions of precious stone deposits in the U.S.S.R., his classification of precious and building stones, and his articles on the part of precious stones in the history of culture, are all classics of our mineralogic literature.

The second group of problems within the scope of A.Ye. Fersman's activity were those of the alteration and formation of minerals on the surface, at times almost under our eyes. These phenomena are very important for an understanding of the formation conditions for a number of mineral deposits, in the process of weathering and redeposition. A.Ye. Fersman has provided some of the classic solutions for these problems. He studied the mobile variable equilibria prevailing in these processes; he demonstrated and emphasized the role of colloid solutions, and established the effect of the physical geographic medium on the formation of minerals. His chemical study of magnesium silicates was very advanced for the time (1912); his research in the mineralogy of clays was of great practical value.

A.Ye. Fersman did not limit his activity to a narrow circle of purely scientific pursuits. He worked enthusiastically all his life on the problems of the industrial development and mineral wealth of the remote reaches of the U.S.S.R., actively participating in the solution of important problems of state economy. Along with outstanding theoretical work, his interests embraced the organization of mineralogic and geochemical work in the newly created institutes and laboratories, a call for creative participation to youth and their education in the true spirit of advanced science, and an incisive analysis of the problems confronting Soviet mineralogy at different stages of its development.

In conclusion, B.A. Fedorovich, Doctor of Geographic Sciences, spoke on "A.Ye. Fersman, a Geographer."

After the Great October socialist revolution, during the reconstruction period, A.Ye. Fersman, with his typical energy, went on with geographic studies. With all his achievements as a mineralogist, geochemist, and crystallographer, geography was not merely his avocation, a "side line." In the diversified activity of Alexander Yevgen'yevich, geography was a traveling companion of geochemistry, because geographic environment is the subject of both.

As a geographer of stature, A.Ye. Fersman became known in this field not only because of the practical results of his explorations in the North (Kola Peninsula), East (Central Asia) and South (Cheleken Island), but also because of his broad organizational activity in the development of geographic science. He initiated the creation of the Commission for the study of productive potential of the country (K.E.P.S.); later on (1925), he worked with and then directed a special committee for the study of the federated and autonomous republics (subsequently, the Commission for Expeditionary Research; now Council for the study of productive forces, S.O.P.S.). During the Great Patriotic War, he organized the Commission for Geologic and Geographic Service to the Red Army. The Institute of the North (now Arctic Institute); Institute of Arctic Geology, now Geography Institute, Academy of Sciences, U.S.S.R.; and the Institute of Aerial Survey, Geodesy, and Cartography — they all came into being by the efforts of Alexander Yevgen'evich to develop geography as a science.

In the words of the speaker, "In remembering today that inspired student of nature and champion of the reconstruction of our economy, it should not be forgotten that the secret of his inexhaustible energy and his inspiring influence on those close to him and on those familiar with his books, was his firm faith in humanity.

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Meetings commemorating the 75th anniversary of the birth of Academician A.Ye. Fersman were held by many other organizations: in Moscow, by the Moscow Society of Nature Students and Moscow Geologic-Prospecting Institute; in Leningrad, by the Mining Institute and the University; in Kiev, by the Academy of Sciences, Ukr.S.S.R. and the Ukrainian Affiliates of the All-Union Mineralogical, Geographic, and Chemical Societies; in Tiflis, by the Georgian Geologic Society and the Society for the Propagation of Political and Scientific Knowledge; in Sverdlovsk, by the Uralian Division of the All-Union Mineralogical Society and the Sverdlovsk Society of Nature Students.

Lectures and talks on the life and activity of A.Ye. Fersman were also given in Kirghizia, the Crimea, and elsewhere.

THE ALL-UNION METALLOGENIC CONFERENCE IN ALMA-ATA[2]

A conference on metallogenic mapping was held in Alma-Ata, December 8-12, 1958.

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[2]O vsesoyuznoy metallogenicheskoy konferentsii v g. alma-ate.

More than 800 persons attended, among them 280 representatives of other republics and cities: Georgia, Armenia, Azerbaidzan, Ukraine, Siberia, Urals, etc. They were members of scientific institutions, the Academy of Sciences, U.S.S.R.; Siberian division of the Academy, republican academies of science, various ministries and departments, local geological administrations, and Sovnarkhoz geologists.

Many papers were read, with tens of persons participating in the discussions or presenting their papers for publication. The meeting took place in a new and beautiful conference hall of the newly completed building of the Academy of Sciences, Kazakh. S.S.R. The Conference had been well organized and ran smoothly. The discussion was facilitated by a preliminary publication of the papers or their abstracts. Of great help was a map display, next to the conference hall.

The conference took place in a businesslike atmosphere and provided very valuable material for a further development of metallogenic ideas and methods.

The conference was opened by an address of P.Ya. Antropov, Minister of Geology. P. M. Tatarinov spoke of the metallogenic research by V.S.E.G.E.I. (All-Union Geological Institute); and K.I. Satpayev, on the comprehnsive exploration maps for central Kazakhstan. He described the now proven methods of mapping as adopted by the Geological Institute, Academy of Sciences, Kaz. S.S.R. Because of this work, the differentiation of geologic formations into six age sequences was established (pre-Paleozoic, early Caledonian, late Caledonian, early Variscan, late Variscan, Kimmeridgian-Alpine). An important phase of the map work was the compilation of all data on ore shows and on the deposit outcrops, with their genetic differentiation into fifty mineralogic formations; also the material of metallometric and concentrate analyses, geophysical anomalies, etc.

A feature of central Kazakhstan, as noted by K.I. Satpayev and other Kazakh geologists, is the association of the mineralization with major regional faults which cut various basement and structural-facies zones, and which control the distribution of ore deposits.

This pehnomenon was emphasized by G.N. Shcherba who described the distribution of rare-metal deposits, and by G.B. Zhilinskiy in his paper on the distribution of tin deposits.

As demonstrated by D.N. Kazanli, geophysical studies establish a deep penetration of fault zones, some of which cut through the entire earth's crust. According to him, these deep-seated fault zones had been initiated a long time ago. They determined the boundaries of downwarps as early as the geosynclinal period. They were repeatedly rejuvenated in the period of platform development. The Kazakhstan example shows that such work is successful only if carried out cooperatively, with the participation of geologists from various departments and organizations.

Much attention was paid to K.I. Satpayev's paper on copper deposits of Kazakhstan.

High praise for the practical results of exploration from maps was the subject of Ts.M. Fishman's talk. He reported that with the help of the maps, the Geological Administration has been successfully carrying on its prospecting work.

General approval was given to the maps of Central Asia. A metallogenic map of southern Tien Shan was demonstrated by Ye.D. Karpova (V.S.E.G.E.I.). This map, built on V.S.E.G.E.I. principles, reflects new elements, as well. Specifically, Ye.D. Karpova separates various structural-facies complexes — those of anticlinal uplifts, interior and foredeeps, superposed troughs, etc. Such a construction affords a diversified approach to the metallogeny of different facies zones, not only on the basis of the data from outcrops, but from the (overall) character of their development. Specifically, the speaker designated four principal structural-facies zones for south Tien Shan: the Fergana-Kokshal and Chatkal-Naryn (zones of downwarping), and Kuramin and Hissar (zones of uplift), all characterized by diverse metallogenic features.

The main Fergana-Kokshal downwarp zone, which underwent a long subsidence, is marked by a development of tungsten deposits, and the Kuraminskiy anticlinal uplift and Chatkal-Naryn foredeep are characterized chiefly by polymetal deposits, commonly associated with limestones. According to Ye.D. Karpova, a belt of ultrabasics and basics passes along the periphery of the Fergana-Kokshal zone. Its rocks are associated not only with early stages, as formerly understood in the V.S.E.G.E.I. sections, but with the main folding stages as well. Thus, they reappear along some of the deep faults. It also is of interest that the same fault zones show an antimony-mercury mineralization, of a later date and proceeding along the same systems of deep faults. The deposits themselves are localized in lithologically and structurally favorable beds, most commonly among limestones overlain by impermeable shales.

An interesting paper on the compilation of more detailed metallogenic maps of the

Chatkal-Kuraminskiy Mountains was given by T.M. Matsokina and written in cooperation with Kh.M. Abdullayev, K.M. Kalabina and a group of geologists.

Besides the Chatkal and Kuraminskiy zones, corresponding to the structural zones of Ye. D. Karpova, the authors note a transitional-boundary fault zone associated with gold and fluorite mineralization. These structural-metallogenic zones are characterized by a different history of development. They also differ in their geology and metallogeny.

Among the graphic material on display, on the Chatkal-Kuraminskiy region, there are interesting paleogeographic and lithologic maps showing the thickness of sediments, the distribution of dikes, and the clearly defined long and narrow belts of dikes along the faults. Very interesting are maps showing the depth of erosion; these maps make it possible to attack the problem of the formation depth for the deposits themselves. As a generalization of all metallogenic and metallogenic-analysis maps, there are geologic exploration maps showing with appropriate symbols the distribution areas for different types of mineralization.

These maps by the Central Asian group have received general approval.

Equally interesting were maps of the Caucasus, particularly those presented by G.A. Tvalchrelidze. He has subdivided the polymetal province of the Caucasus into structural-facies zones of various ages and structural positions. Shown within these zones are sedimentary, igneous, and ore formations — prefolding, folded, and postfolding. This is a modification of Yu.A. Bilibin's idea as applied to a complex ore province, with non-contemporaneous cycles of igneous activity and mineralization.

A paper on the metallogenic map of Armenia was read by S.S. Mkrtchan.

Papers on the metallogenic maps of the Urals (M.M. Aleshin, V.P. Pervov, N.V. Kuklin) demonstrated the character of mineralization chiefly as a function of the igneous complexes. The latter are distributed here lineally, which has determined the linear distribution of ore belts.

P.M. Lazarev and I.V. Lennykh spoke on the patterns in the distribution of pyrite deposits in South Urals. The authors noted the broad age range for the pyrite mineralization and its constant spatial association with albitophyres. Their exploration map is an example of large-scale metallogenic mapping.

A different type of large-scale maps for the area of pyrite ore development was presented for the Altai, by P.F. Ivankin, A.K. Kayupov, and G.N. Shcherba. This region exhibits a close association of pyrite deposits with zones of fracturing and other fault zones.

An important factor in the localization of ore deposits is the presence of a sedimentary volcanic complex, where the most favorable ore-bearing structures are represented by faults and fault zones; also by folds, especially domal, in sharply anisotropic media, etc.

Of a special type is a metallogenic map of the Northeast, reported by V.T. Matveyenko. This map has been constructed on a tectonic basis, after N.S. Shatskiy's principle of the making of tectonic maps. It shows the individual complexes of principal structural stages, with the relative age of enlarged age complexes designated by coloring.

The platform-type ore regions were described by N.P. Semenenko, Yu.G. Staritskiy, and others; also by M.N. Godlevskiy.

In his paper, "Metallogenic Epochs, and Exploration Maps of Ore Deposits in the Ukrainian S.S.R.," N.P. Semenenko emphasized the need for more attention to be paid to platforms, inasmuch as they contain large ore deposits and are even more important for the world reserves of ore than the younger geosynclinal provinces. The speaker stressed the fact that patterns in the localization of ore deposits, in ancient folded provinces (Precambrian), are the same as for folded provinces with a younger metallogeny. On the other hand, the mineralizing significance of the Precambrian is very great, because that period is many times longer than the historic geologic period. In its turn, it can be differentiated into several epochs commensurable with later epochs of the Caledonian, Variscan, and other tectonic-igneous cycles.

The map shows the most important regularities in the distribution of ore deposits, in both the Precambrian basement and the platform mantle; and especially of placer deposits, the ancient weathering crust, and formations related to faults in the ancient crystalline complex.

Specially designated on the map are younger ore provinces: the Donets folded zone and the still younger Carpathian and coastal Crimean foldings.

The problem of platforms was considered in a collective paper of Yu.G. Staritskiy, V.L. Masaytis, V.I. Dragunov, and N.S. Malich. The authors are concerned chiefly with the platform mantle; they note that ore deposits in the crystalline basement should

be studied, on the whole, according to the methods applied to geosynclinal zones. Regarding the development of the platform mantle, they emphasize the relationship between tectonic elements of the platform and of the adjacent geosynclinal provinces. They noted certain patterns in the development of sediments in superposed syneclises, as well as those in the distribution of ore deposits associated with platform faults.

This aspect of the problem was considered in particular detail, in a paper of M.N. Godlevskiy who described the principles and methods of metallogenic mapping for the Noril'sk region. He noted the representation of major faults along the edges of the Tungus syneclise, which are associated with trap intrusions. Of great importance in metallogenic investigations is the study of petrography of trap massifs and the discovery in them of the most favorable differentiated varieties distinguished by quite definite petrochemical and mineralogic features.

V.A. Kuznetsov spoke on the metallogeny connected with basic and ultrabasic magmas of western Siberia.

Of especial interest was a paper by B.M. Gimmel'farb (Himmelfarb) on the making of exploration maps in some of the areas of the Turgai depression.

Papers of V.I. Smirnov, Ye.T. Shatalov, Ye.A. Radkevich, and others, dealt with general problems in methodology.

V.I. Smirnov exhibited an attempt at a metallogenic regional differentiation of the Soviet Union, on the basis of the distribution of ore deposits of various ages. His map shows the provinces of polycyclic mineralization, with superposition of noncontemporaneous ore shows.

The paper of Ye.T. Shatalov and A.V. Orlova described a method of large-scale metallogenic mapping. Papers of Ye.A. Radkevich and I.N. Tomson dealt with the same subject. The speakers emphasized the importance of scaled representations for the results of study, and the necessity of a special approach to that task.

In her paper, "Types of Ore Areas," Ye. A. Radkevich stressed the necessity of a more diversified approach to different structures. She pointed out that the mineralization types depend not so much on the time of manifestation of this or that stage as on features of the development and structural position of various structural-tectonic zones, such as interior troughs, uplifts, edge faults, etc. Two types of zones have been separated: a femic, with basic igneous activity along deep faults; and a sial, with the preponderance of granites.

The problem of classification of ore deposits was also touched upon in papers by V.A. Nikolayev, Kh.M. Abdullayev, G.A. Sokolov, and others.

There were lively discussions on the papers, and talks on the results of metallogenic research. Many speakers underscored the equivalence of different approaches and methods, and the necessity of a diversification in the carrying on with the research, in order not to lapse into a rut (V.A. Nikolayev, Kh.M. Abdullayev, D.I. Shcherbakov, G.A. Sokolov, and others). It was stated, in effect, that different methods and techniques should be used for different purposes, for different map scales, and different terrains. In addition, there were voices expressing the necessity of an overall view and of general instructions for different types of metallogenic research. Many speakers pointed out the necessity of long-range exploration which would be carried out on the basis of progressive study.

The resolutions contain basic organizational measures for the development of metallogenic research and of its various trends and methods. Specifically, it was pointed out that the success of such research would depend to a considerable degree on the extent of participation by the geologists of local industrial organizations. It was emphasized that the geologic administrations should take part in the metallogenic work and should include it in their schedule.

It has been agreed that maps should be divided into 1) index (1:1,000,000 and smaller); 2) intermediate scale (1:1,000,000 to 1:200,000); 3) large scale (1:200,000 and larger).

The first type is for the representation of most general regularities and for the planning of small-scale work. The second and third are for servicing the needs of local geologic administrations, and for the planning and direction of exploration and prospecting. The importance of large-scale metallogenic research is emphasized. It should include the making of a whole series of maps, from situation maps to the metallogenic-exploration, indicating the order of work to be planned.

The Conference adopted a resolution on the publication of maps and the accompanying text in large editions, to be used as manuals. An exchange of opinions is also contemplated, in conferences held by the Commission on patterns in the distribution of useful minerals, and by their local divisions.